Bulletin
of
The Geological Society of America

VOLUME 68 November 1957 NUMBER 11

CONTENTS

	Pages
Pebble and sand lithology of the major Wisconsin glacial lobes of the Central Lowland. By Richard C. Anderson	1415–1450
Correlation of the Triassic formations of North America exclusive of Canada. By the Triassic Subcommittee, John B. Reeside, Jr., *Chairman, et al.*	1451–1514
Thermal effects of the ocean on permafrost. By Arthur H. Lachenbruch	1515–1530
Structure and petrology of the Caribou Lake intrusive body, Ontario, Canada. By Gerald M. Friedman	1531–1564
Short Notes	
Use of geology in planning the Normandy invasion. By Charles T. Snyder	1565–1566
Additional notes on use of geologists in the European Theater of Operations during World War II. By Louis J. Simon	1567–1568
Supposed Permian tillites in northern Mexico are submarine slide deposits. By Norman D. Newell	1569–1576
Lowerre quartzite problem. By Matthew F. Norton and Ross F. Giese, Jr.	1577–1580
Polymorphism of micas in the Mineral Bluff and Epworth quadrangles, Georgia. By Vernon J. Hurst	1581–1584

Subscription $15.00 per year.

Printing Office: Mt. Royal & Guilford Aves., Baltimore 2, Md.

Communications for publications should be addressed to The Geological Society of America, Dr. H. R. Aldrich, Secretary, 419 West 117 Street, New York, N. Y.

Please give notice of change of address 4 weeks in advance. Claims for nonreceipt of the Bulletin of a given month must be sent to the Secretary of the Society before the end of the second succeeding month.

Entered as second-class matter at the Post Office at Baltimore, Md., under the Act of Congress of July 16, 1894.
Accepted for mailing at special rate of postage provided for in Section 1103 Act of October 3, 1917, authorized on July 8, 1918.

PAPERS IN PRESS FOR FORTHCOMING ISSUES

HALOGEN-ACID ALTERATION OF ASH AT FUMAROLE NO. 1, VALLEY OF TEN THOUSAND SMOKES, ALASKA. By T. S. Lovering

ORIGIN OF THE CORUNDUM DEPOSITS OF RENFREW COUNTY, ONTARIO, CANADA. By H. D. Carlson

CORONITES FROM INDIA AND THEIR BEARING ON THE ORIGIN OF CORONAS. By M. V. N. Murthy

STRATIGRAPHY AND STRUCTURE OF THE ST. ALBANS AREA, NORTHWESTERN VERMONT. By A. B. Shaw

THERMAL WATERS OF VOLCANIC ORIGIN. By D. E. White

MAGNETIC, CONNATE, AND METAMORPHIC WATERS. By D. E. White

RECENT UNDERWATER SURVEYS USING LOW-FREQUENCY SOUND TO LOCATE SHALLOW BEDROCK. By W. O. Smith

GEOLOGY OF THE MT. GARIBALDI MAP-AREA, SOUTHWESTERN BRITISH COLUMBIA, PARTS I AND II. By W. H. Mathews

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BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1415–1450, 2 FIGS. NOVEMBER 1957

PEBBLE AND SAND LITHOLOGY OF THE MAJOR WISCONSIN GLACIAL LOBES OF THE CENTRAL LOWLAND

BY RICHARD C. ANDERSON

ABSTRACT

Pebble and sand-grain counts were taken from the axial portions of moraines of six Wisconsin glacial lobes of the Central Lowland (Erie, Saginaw, Lake Michigan, Green Bay, Des Moines, and Iowan), and the observed distribution of lithologic types was interpreted in terms of provenance, lithologic properties, and glacial processes. It appears unlikely that the glacial lobes of the Central Lowland can be differentiated on the basis of a single "indicator" lithology. Although several lithologies were found to be distinctive of some lobes and not of others, only the Saginaw lobe contained lithologies not found elsewhere (jasper conglomerate and a Precambrian conglomerate resembling tillite). Investigation of the relative proportion of lithologic types appears to be the most fruitful method for the differentiation of glacial lobes.

Several pebble counts taken from Kansan till in eastern Iowa and western Illinois substantiate this general method by showing strong lithologic affinities toward the Iowan drift and thus indicating a Keewatin source. This conclusion has been reached by others using various methods of approach.

CONTENTS

TEXT

	Page
Introduction	1415
Acknowledgments	1416
Methods of study	1417
Sampling	1417
Laboratory procedure	1417
Pebble lithology of the lobes	1418
Precambrian fraction	1418
Total pebble fraction	1421
Sand lithology of the lobes	1424
Lithologic homogeneity within lobes	1425
Interpretation of drift lithology	1426
Pebble counts from Kansan till	1430
Conclusions	1431
References cited	1431

ILLUSTRATIONS

Figure	Page
1. Location of sand and pebble samples	1416
2. Average pebble and sand lithology by lobe	1419

TABLES

Table	Page
1. Comparison of ratios between sand and pebble fractions	1424
2. Counts of Precambrian pebbles from the Erie lobe	1432
3. Counts of Precambrian pebbles from the Saginaw lobe	1433
4. Counts of Precambrian pebbles from the Lake Michigan lobe	1434
5. Counts of Precambrian pebbles from the Green Bay lobe	1435
6. Counts of Precambrian pebbles from the Des Moines lobe	1436
7. Counts of Precambrian pebbles from the Iowan drift	1437
8. Pebble counts from the Erie lobe	1438
9. Pebble counts from the Saginaw lobe	1439
10. Pebble counts from the Lake Michigan lobe	1440
11. Pebble counts from the Green Bay lobe	1440
12. Pebble counts from the Des Moines lobe	1441
13. Pebble counts from the Iowan drift	1442
14. Sand-grain counts from the Erie lobe	1442
15. Sand-grain counts from the Saginaw lobe	1443
16. Sand-grain counts from the Lake Michigan and Green Bay lobes	1443
17. Sand-grain counts from the Des Moines lobe and Iowan drift	1444
18. Pebble counts from Kansas till	1444
19. Counts of Precambrian pebbles from Kansan till	1445
20. Description of sample locations	1446

INTRODUCTION

PURPOSE OF STUDY: The composition and texture of unweathered glacial till is a reflection of the materials over which the ice has passed. Hence, the lithologic constituents of a moraine must be dependent upon the source of the ice and the composition of its floor. Comparison of the composition of till deposited by ice moving

over different bedrock lithologies should therefore show differences intimately related to the path that the ice followed.

The present study is an attempt to interpret the pebble and sand lithology of six Wisconsin glacial lobes in terms of provenance, lithologic properties, and glacial processes; to establish for each lobe lithologic indices indicative of the source of the ice and the path of its advance; and to evaluate this procedure as a method of differentiating till sheets. The pebble and sand-grain counts and the location of samples are shown in Tables 2–20 which are assembled at the end of the paper.

REGIONAL SETTING: The following Wisconsin glacial lobes were sampled because they probably represent the major lines of ice movement into the Central Lowland throughout the Pleistocene: Erie, Saginaw, Lake Michigan, Green Bay, Des Moines, and Iowan (Fig. 1). These lobes should be regarded as separate flow units of a continuous ice sheet, which were initiated and controlled by the bedrock topography (Horberg and Anderson, 1956, p. 112).

To obtain a representative sample of the pebble and sand lithology of the till of each lobe, only moraines that show a clear relation to the parent lobe were sampled. Thus the samples from the Erie and Lake Michigan lobes were taken from the Cary moraines, since in each case they indicate without question the path of the ice that deposited them. In the Saginaw, Green Bay, and Des Moines lobes both Cary and Mankato moraines were sampled because they show unequivocal affinities to the parent lobe. Iowan drift was sampled only in eastern Iowa.

PREVIOUS STUDIES: Studies of drift-pebble lithology have been conducted by a number of workers over widely spaced areas, and the results, generally bearing on local correlation problems, are to be found largely in regional reports. A few, however, merit special mention because they deal with problems similar to those encountered in the present study. Lundquist (1935) investigated the pebble composition of certain drift sheets in Sweden and presented the data by a variety of graphical methods. Milthers (1942) differentiated drifts in Denmark on the basis of distinctive Norwegian and Baltic lithologic types. MacClintock (1933) established a western source for Nebraskan till and an eastern source for Kansan till in south-central Illinois by the use of pebble counts. Holmes (1952) made a comprehensive study of drift lithology in west-central New York and found relationships between the distribution of certain rock types and the bedrock topography. Dreimanis and Reavely (1953) made comparative studies, including pebble counts, on two tills along the north shore of Lake Erie; they recognized definite lithologic differences in the pebble-size grade. Horberg (1956) used pebble counts as an aid in distinguishing the Illinoian from the older drifts in western Illinois.

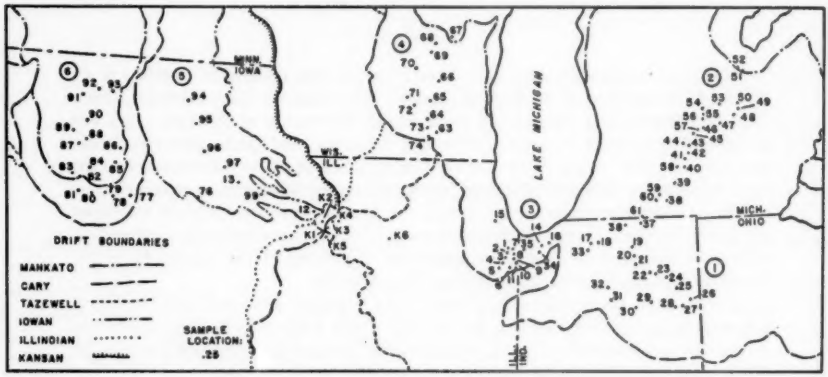

FIGURE 1.—LOCATION OF SAND AND PEBBLE SAMPLES

Circled numbers: 1—Erie lobe, 2—Saginaw lobe, 3—Lake Michigan lobe, 4—Green Bay lobe, 5—Iowan drift, 6—Des Moines lobe (after Flint *et. al.*, 1945; Bretz, 1951; Ruhe, 1952; Horberg and Anderson, 1956).

ACKNOWLEDGMENTS

The writer wishes to record his thanks to the late Leland Horberg of the University of

Chicago for helpful advice and criticism throughout the course of this study, and to D. J. Fisher, M. M. Leighton, F. W. Thwaites, and J. H. Zumberge for critically reading the manuscript.

Methods of Study

Sampling

LOCATION: Sampling was designed to yield data giving an accurate picture of the pebble and sand lithologies carried by ice advancing along well-defined paths from known areas of accumulation. In addition it was desired that the samples be collected in a manner permitting valid comparisons to be made between lobes and that they be as representative of a given lobe as possible. Therefore two traverses were made across each lobe parallel to the direction of ice movement, as close as possible to the axis of the lobe, and across only those moraines whose relations to the parent lobe are unquestioned. Aside from those of the Iowan and a few from the Green Bay lobe, all the samples were taken from end moraines. In the Green Bay lobe it was necessary to take three samples from the drumlin field of southeastern Wisconsin, which is largely devoid of moraines. Because the Iowan drift has no end moraines, the samples were collected at about 20-mile intervals.

MATERIAL SAMPLED: The samples were taken from till wherever possible. In two areas sandy or gravelly kame moraine was sampled: the La Grande, Sturgis, and Kalamazoo moraines of the Saginaw lobe; and the Johnstown, Milton, and Lake Mills moraines of the Green Bay lobe. Three of the Iowan drift samples are from outwash.

SAMPLE SIZE: Three samples were collected from each location: a count of 100 pebbles, a count of 100 Precambrian pebbles, and about 1 kg of till from which the sand fraction was later extracted for lithologic examination. Experience has shown that samples of 100 pebbles are sufficiently large to give accurate pictures of the relative abundance of the locally derived constituents and their percentage of the total. The far-travelled, Precambrian lithologies, however, are most consistent within a given lobe, most indicative of a particular source, and hence most valuable for differentiating lobes. Since the number of Precambrian pebbles usually obtained in a count of 100 pebbles is not large enough to establish the relative amounts of the various Precambrian lithologies, an additional sample of 100 Precambrian pebbles was collected.

INDEX GRADE SIZE: Each grade size in glacial till has a unique distribution of lithologic types. The size which shows the greatest variety of lithologies affords the best possibilities for the differentiation of drift sheets. Because the ½- to 1-inch-size grade contains the greatest variety of rock types (Horberg and Potter, 1955, Fig. 3), all the counts were taken from this range.

PROCEDURE IN THE FIELD: Pebble samples were taken from the surface of till exposures in horizon 4 or 5 of the profile of weathering (Leighton and MacClintock, 1930). First the total pebble sample was collected from a unit area of the surface; all pebbles of index grade size were removed from one area before going on to the next. The Precambrian pebble sample was then taken in the same manner. The till samples were also taken from horizon 4 or 5 of the weathering profile and far enough below the surface of the exposure to eliminate contamination by slope wash.

Laboratory Procedure

In the laboratory the pebbles were washed, identified, and counted, and an index collection of typical and diagnostic lithologies was compiled for each lobe. The till samples were air-dried and broken into lumps about 1 inch in diameter with a mortar and pestle. Several of these lumps (300–500 gms), which constituted the laboratory sample, were soaked for about an hour in a solution of 0.25 gm of Calgon per liter of water until they were disaggregated. The sample was then washed through a .124 mm screen, and the finer fraction was discarded. The coarser fraction was separated into two fractions by dry sieving: .124–.500 mm and .500–.991 mm. The coarser of these two fractions was coned and quartered until reduced to 300–400 grains which were examined under the binocular microscope. One hundred sand grains were counted by removing them with tweezers from the periphery of the microscope sample. The results of several counts of various sizes up to the whole microscope sample and the consistency of the counts of 100 grains within a single lobe (Tables 14–17) indicate that a count of 100 grains is sufficiently accurate to corroborate the pebble-count data. Except for distinguishing quartz from feldspar and limestone from dolomite, the lithology of the .500–.991 mm grade size can readily be determined. The limestones and dolomites were grouped to-

gether under the microscope and then removed and tested in a dilute solution of HCl in which limestone effervesced vigorously and dolomite evolved only a small train of bubbles. Quartz and feldspar were grouped together with minor amounts of other similar minerals. In addition, the quartz-feldspar group was subdivided into rounded and angular grains; a grain was designated as rounded if its visually estimated roundness index exceeded 0.6 (Pettijohn, 1949, p. 49–52).

Pebble Lithology of the Lobes

Precambrian Fraction

ERIE LOBE: Igneous rocks constitute 69.5% of the Precambrian pebbles of the Erie lobe metamorphic rocks 25.1%, and vein quartz and other minor lithologies 5.4% (Table 2). Whitish-pink to red Laurentian granites are common, but the most abundant granitic type is a dark-red, fine-grained, hybrid granite which appears to grade into a true gneiss. It forms 11.0% of the Precambrian as compared with 8.3% for the typical Laurentian granites. The remainder of the igneous rocks can be subdivided into medium-basic and basic types. The former are generally coarse-grained diorites which are commonly weathered brown on the surface. They constitute 17.8% of the Precambrian. The basic pebbles have been further subdivided on the basis of grain size. Most of the coarse-grained varieties are gabbro, whereas most of the fine-grained types are basalt. Basic pebbles form 32.4% of the Precambrian. The Erie lobe has a higher percentage of coarse-grained basic types (8.4) than any other lobe and is exceeded only by the Lake Michigan and Green Bay lobes in the amount of basaltic types (24.0%). Banded basalt pebbles with denser, more resistant layers standing in relief are common (1.8%). Greenstone pebbles have not been included with the basic types, although every gradation to a basalt can be recognized. Only those pebbles that were unmistakably green were designated as greenstone, and in the Erie lobe they form 4.4% of the Precambrian.

Medium-grained quartzite, which varies from red to gray, is the most abundant metamorphic lithology (10.0%). Graywacke ranks next among the metamorphic types with 5.9%. Included with the graywackes are dark, dense pebbles with detrital quartz grains as well as the more easily recognized coarser-grained, dark metasediments. Gneiss forms 4.9% of the Precambrian, and black slate with well-developed cleavage is found in minor amounts (1.5%). Hornblende schist makes up 2.8% of the Precambrian pebbles. Other minor lithologies include vein quartz, 0.9%, and rhyolite 0.1%.

SAGINAW LOBE: The principal Precambrian lithologies of the Saginaw lobe are similar to those of the Erie lobe, but they are found in different proportions. Igneous rocks constitute 63.5% of the Precambrian pebbles of the Saginaw lobe, metamorphic rocks 27.6%, and greenstone, vein quartz, and other minor lithologies 9.0% (Table 3). White to pink Laurentian granites are the most abundant granitic type; they make up 14.9% of the Precambrian, whereas the dark red, fine-grained, gneissic granites which are common in the Erie lobe make up only 2.2%.

Medium-basis pebbles are more common in the Saginaw than in the Erie lobe; they make up 22.6% of the Precambrian (Fig. 2). Basic pebbles decrease from 32.4% to 23.8%. Of this, 4.5% is coarse-grained; this is only about half as much as in the Erie lobe. Greenstone forms 6.6% of the Precambrian.

As a result of an influx of quartzite pebbles (14.8%), the Saginaw lobe has more metamorphic pebbles than any other lobe studied. They are lithologically similar to those of the Erie lobe. The 6.3% graywacke is nearly the average amount found in the four lobes east of the Mississippi River, but it is considerably less than that of the Des Moines and Iowan drift sheets. Schist forms 2.9% of the Precambrian, which is almost the same as in the Erie lobe. This includes 0.5% green schist, which was not found in the Erie till. Gneiss and slate are less abundant than in the Erie lobe and constitute 3.3 and 0.3% of the Precambrian, respectively.

In addition to green schist several other Precambrian lithologies were found in the Saginaw but not in the Erie lobe. Red sandstones and siltstones of the Keweenawan (Lake Superior) type (0.2%), jasper and jasper conglomerate (0.4%), and a slightly metamorphosed conglomerate having a fine-grained matrix and a wholly unsorted aspect resembling tillite (0.2%) were not found in the till of the Erie lobe. The tillite (?) and jasper conglomerate appeared only in Saginaw till. Slosson (1933), however, found jasper conglomerate distributed in Illinoian till from Ohio to southeastern Iowa.

Vein quartz and rhyolite are found in minor amounts, 1.2 and 0.4%, respectively.

LAKE MICHIGAN LOBE: Igneous rocks constitute 75.3% of the Precambrian pebbles of the Lake Michigan lobe; metamorphic rocks

18.0%; greenstone, vein quartz, and other minor lithologies 6.8% (Table 4). All the granite pebbles are of the typical coarse, red

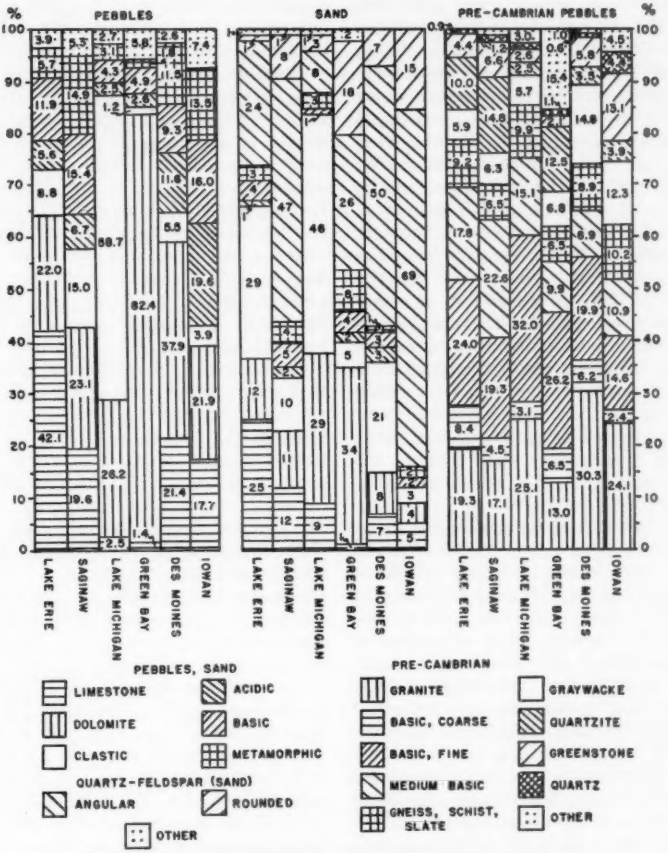

FIGURE 2.—AVERAGE PEBBLE AND SAND LITHOLOGY BY LOBE

others in the amount of metamorphic pebbles. This is the more remarkable because it contains the highest percentage of schist (8.0), of which to white Laurentian varieties. The amount of granite (25.1%) is greater than any lobe east of the Mississippi River and is exceeded only by the Des Moines lobe. More basic pebbles were found in Lake Michigan till than in that of any other lobe, but the medium-basic types dropped to 15.1%, compared with 22.6 and 17.8%, respectively, in the Saginaw and Erie lobes (Fig. 2). Coarse-grained basic types show a slight decrease to 3.1%, but fine-grained basic types are abundant. The 2.6% greenstone is less than in either the Saginaw or Erie lobes.

The Lake Michigan lobe ranks below all others in the amount of metamorphic pebbles. 0.7% is a green variety similar to that found in the Saginaw lobe. The other metamorphic lithologies are neither distinctive nor found in as great amounts as elsewhere. The percentages of graywacke (5.7) and quartzite (2.5) rank lowest among the lobes, whereas gneiss (1.9%) is only slightly more abundant than in the Des Moines lobe, which ranks lowest in this respect.

Clastics of the Keweenawan type increase from 0.2% in the Saginaw lobe to 1.7% in the Lake Michigan, and jasper also increases from a trace to 0.4%. Jasper conglomerate and

tillite(?) are absent. Rhyolite (0.8%) was found to be more abundant than in the Saginaw lobe. Vein quartz (1.2%) and rhyolite porphyry (0.1%) also occur.

GREEN BAY LOBE: Igneous rocks constitute 55.6% of the Precambrian pebbles of the Green Bay lobe; metamorphic rocks 25.8%; Keweenawan clastics 8.3%; and greenstone, rhyolite, rhyolite porphyry, and other minor lithologies 10.3% (Table 5). Igneous rocks are considerably less abundant than in the other three lobes east of the Mississippi River (Fig. 2). A great part of this difference can be attributed to a reduction of the granite to 13.0%. The majority of the granite pebbles are coarse granites of the Laurentian type, but some fine-grained, dark-red, gneissic varieties are present (1.7%). A comparable reduction in the amount of medium-basic pebbles to 9.9% also accounts for part of the decrease in the amount of igneous rocks. The percentage of coarse-grained basic types (6.5) is somewhat higher than in the Lake Michigan and Saginaw lobes. The percentage of fine-grained basic types (25.4) is comparable with those of the lobes to the east, although it is lower than that of the Lake Michigan lobe.

Quartzite is the most common of the metamorphic lithologies of the Green Bay lobe and is also the most inconsistent from sample to sample because of local Huronian sources. It forms 12.5% of the Precambrian, but if sample 72 (Table 5), having 67% quartzite, is omitted, the average over the lobe as a whole is reduced to 7.0%. The percentages of graywacke (6.8) and gneiss (4.5) are somewhat higher than in the Saginaw and Lake Michigan lobes but are nearly the same as in the Erie lobe. In contrast to the adjoining Lake Michigan lobe, having 8.0% schist, only 2.0% schist was found in the Green Bay lobe.

Relatively large amounts of Keweenawan clastics (8.3%), jasper (1.6%), rhyolite (2.2%), rhyolite porphyry (2.6%), and iron formation (0.7%) are diagnostic. Generally they are abundant enough to give the pebble fraction of the till a reddish hue, notwithstanding the large amounts of dolomite present.

DES MOINES LOBE: Igneous rocks constitute 65.3% of the Precambrian of the Des Moines lobe, metamorphic rocks 27.2%, and greenstone and other minor lithologies 7.4% (Table 6). The Des Moines lobe has 30.3% granite, substantially more than any other lobe (Fig. 2). All the granite is essentially Laurentian in type. Medium-basic pebbles form only 8.9% of the Precambrian, and they appear on the whole to be finer-grained than those from the eastern lobes. Coarse-grained (6.2%) and fine-grained (19.9%) basic types are found in about the same amount as elsewhere. The 5.8% greenstone in the Des Moines lobe is exceeded only by the Saginaw among the eastern lobes but is far below the 13.1% in the Iowan.

The Des Moines lobe contains significantly more graywacke (14.8%) than the eastern lobes. It is also high in schist (7.6%) and ranks second to the Lake Michigan lobe in that respect. Hornblende (7.4%) and green (0.2%) schists are present. The Des Moines lobe is low in gneiss (1.1%) and quartzite (3.5%), but the lithologies are essentially the same as elsewhere. A red to maroon variety of quartzite, forming 2.1% of the Precambrian, may represent the Sioux Quartzite of southwestern Minnesota. Black slate makes up 0.2% of the Precambrian of the Des Moines lobe.

Keweenawan red clastics are absent from the Des Moines lobe, but a red argillite, thought to be derived from the Sioux Quartzite, forms 0.1% of the Precambrian. Rhyolite, rhyolite porphyry, and quartz are present in small amounts, 0.5, 0.4, and 0.6%, respectively.

IOWAN DRIFT: Igneous rocks constitute 51.9% of the Precambrian pebbles of the Iowan drift; metamorphic rocks 26.2%; greenstone 13.1%; and vein quartz and other minor lithologies 8.9% (Table 7). The Iowan drift contains 24.1% granite, slightly more than the average of all the lobes, and it is almost without exception the whitish-pink to red Laurentian type. The Iowan drift is relatively impoverished in the basic igneous lithologies. The medium-basic and basic types form only 27.8% of the Precambrian, compared with an average of 47% for the other five lobes (Fig. 2). As in the Des Moines lobe, the medium-basic pebbles (10.9%) are finer-grained than those found in the eastern lobes. More medium basic pebbles were found in the Iowan than in the Des Moines, but the amount of basaltic and gabbroic types is very low in the Iowan, 14.6 and 2.4%, respectively. Greenstone is abundant in the Iowan drift (13.1%); its percentage is twice as great as that of the Saginaw lobe which has the second highest amount.

Metamorphic pebbles are present in the Iowan drift in about the same amounts as in the Des Moines lobe. The 12.3% graywacke is less than in the Des Moines but still well above the average of the eastern lobes. Hornblende schist forms 6.1% of the Precambrian; green schist is also present (0.4%). Compared with the Des Moines lobe both gneiss (2.9%) and

quartzite (3.9%) are more abundant in the Iowan drift. Black slate forms 0.8% of the Precambrian.

Red clastics from the Lake Superior region are found in the Iowan (0.8%) with relatively large amounts of rhyolite and rhyolite porphyry, 1.8 and 1.4%, respectively. The latter is second in abundance only to the Green Bay lobe and is considerably larger than in any of the other lobes. Quartz makes up 4.4% of the Precambrian, about 4 times greater than the average of the other five lobes.

SUMMARY: The general picture which emerges from the study of counts of Precambrian pebbles is one of differences between lobes in the relative proportions of several ubiquitous lithologies rather than of limited occurrences of distinctive lithologies, although the latter are present. Thus, less granite was found in the lobes east of the Mississippi River than in the Des Moines and Iowan to the west. Conversely, the amount of basic pebbles decreased from east to west. The four eastern lobes average close to 6% graywacke, whereas twice as much appears in the two western lobes. Although the Lake Michigan and Green Bay lobes are somewhat low in greenstone (2.6 and 2.1%, respectively), only the Iowan has a notable amount; its 13.1% is about 3 times the average of the other lobes. Schist shows little systematic change, but the Lake Michigan, Iowan, and Des Moines lobes, which have high percentages of schist, also have large amounts of granite and small amounts of quartzite and gneiss. Rhyolite and rhyolite porphyry were found in every lobe, but they form more than 1% of the Precambrian only in the Green Bay and Iowan drifts, where they are 4.8 and 3.2%, respectively. Quartz averages about 1% of the Precambrian, except in the Iowan which has 4.4%.

Three Precambrian lithologies were found to be restricted in occurrence and may serve as useful index lithologies. Pebbles of Huronian tillite(?) and jasper conglomerate occurred only in the till of the Saginaw lobe, and taconitic iron formation from Upper Michigan was found only in the Green Bay lobe.

In addition two other Precambrian lithologies may be accepted as indicating a general provenance region but can be used for differentiating tills only on the basis of their relative abundance. The Keweenawan sandstone and siltstone pebbles point unequivocally to the southern shore of Lake Superior as their source. but they are not confined to one lobe. As could be expected, however, the Green Bay lobe contains the greatest number of these pebbles; the contiguous Lake Michigan lobe ranks second. Jasper-bearing bedrock appears to be localized in Upper Michigan and in an area north of Lake Huron, and its distribution in the till parallels that of the Keweenawan clastics. Neither of these lithologies was found in the Erie or Des Moines lobes. Thus the conclusion that pebbles of jasper and Keweenawan red clastics are characteristic of only the central four of the six lobes under consideration in this study seems justifiable.

Although the Lake Superior lobe in Minnesota was not included in this study, it contains abundant Keweenawan clastics, rhyolite, jasper, and iron formation. During the late Wisconsin the Lake Superior lobe was not extensively developed, but the occurrence of its characteristic lithologies in the Iowan and Kansan in northeastern Iowan suggests that it was considerably more extensive during at least one of the earlier glaciations.

Total Pebble Fraction

ERIE LOBE: The pebble fraction of the till of the Erie lobe contained 24.1% Precambrian and 75.9% Paleozoic lithologies (Fig. 2). Limestone makes up 42.1% of the total, of which 35.8% is brown gray to black, dense, and fossiliferous (Table 8). It is thought to be the Ordovician Trenton and Black River limestones bordering and underlying Lake Ontario. The remainder of the limestone is brown to buff, sandy, and porous. It is not a distinctive lithologic type, and its provenance is uncertain. Its nonresistant character indicates a local source which may be the Devonian of northwestern Ohio and northeastern Indiana.

Buff to gray-black, porous to dense, cherty Silurian dolomite is common in the Erie lobe and makes up 22.0% of the total. Buff to white chert and siliceous coral, closely associated with the dolomite, make up 3.0%.

Devonian clastics constitute 8.1% of the pebbles of the Erie lobe and appear to be largely derived from the Ohio shale in northwestern Ohio and its equivalents in northeastern Indiana. The most abundant lithologies are dark-brown, poorly bedded shale and brown-gray, well-indurated siltstone which together form 7.2% of the total. The most distinctive lithology, however, is gray to black fissle shale (0.9%), which is not as resistant as the other clastic lithologies but is of widespread occurrence. It is derived from exposures of the Upper Devonian New Albany Shale in northern

Indiana, lower Michigan, and northwestern Ohio. A reddish-brown, fine-grained, friable quartz sandstone (0.8%) was found in the Erie lobe. It undoubtedly was derived from a local source, perhaps weathered, iron-stained exposures of Pennsylvanian Sylvania Sandstone in southeastern Michigan or northwestern Ohio.

SAGINAW LOBE: The pebble fraction of the till of the Saginaw lobe contained 39.1% Precambrian and 60.9% Paleozoic lithologies (Fig. 2). Two contrasting limestone lithologies make up 19.6% of the total (Table 9). The most abundant (15.3%) is brownish-gray to black, dense, Ordovician Trenton and Black River limestone similar to that of the Erie lobe. The other is a buff, porous, sandy limestone which is derived from local sources and forms 4.3% of the total. It may represent the Mississippian Bayport Limestone, but there are several possible Carboniferous sources.

The variety of clastic lithologies in the pebble fraction of Saginaw till it is a reflection of the Mississippian and Pennsylvanian bedrock which underlies the entire lobe in lower Michigan. Few of the sandstones, siltstones, and shales are distinctive, and tracing them to a specific source is difficult, if not impossible. Of the clastic Carboniferous lithologies in Saginaw till the most distinctive is a buff-white, fine-grained, well-cemented quartz sandstone (3.6%) which probably is the Mississippian Marshall Sandstone, although it may be derived, at least in part, from the Pennsylvanian Parma Sandstone. The pebbles are similar to the Marshall Sandstone, as seen in exposures in Calhoun County, Michigan. The remaining Carboniferous clastic lithologies consist largely of brown to red, poorly indurated siltstones with minor amounts of reddish-brown to gray shales and ferrugenous concretions. They may have been derived from any of several local sources. Because of their bedrock distribution, Devonian siltstone and black, fissile New Albany Shale were found in large numbers in the Saginaw lobe only in northern Indiana.

Dolomite is common in the Saginaw lobe and makes up 23.1% of the total. Buff to gray, porous to dense, cherty dolomite typical of the Silurian Niagaran Series is less abundant (4.7%) than in the Erie lobe, but a brown, dense variety occurs which is peculiar to the Saginaw lobe and makes up 18.5% of the total. All the dolomite and chert (3.2%) is from the Silurian of the Georgian Bay region.

LAKE MICHIGAN LOBE: The pebble fraction of the till of the Lake Michigan lobe contains 10.7% Precambrian and 89.3% Paleozoic lithologies. Limestone forms only 2.5% of the pebbles of the Cary moraines of the Lake Michigan lobe (Fig. 2). The gray to black dense Ordovician limestones which are so common in the Erie and Saginaw lobes make up only 1.7% of the total in the Lake Michigan lobe, but they are the most abundant limestone type (Table 10). Brown-buff, sandy, porous limestone pebbles similar to those found in the Erie lobe also occur (0.8%). They may be derived from Devonian limestones in the Lake Michigan basin, but no specific source is indicated.

Dolomite and chert, 26.2% and 1.9% respectively, are only slightly more abundant in the Cary moraines of the Lake Michigan lobe than in the Erie and Saginaw lobes. The Marseilles moraine (Tazewell), representative of earlier advances of the Lake Michigan lobe west of the Niagara cuesta, however, consists of 68% dolomite (Anderson, 1955), but these figures are not included in the present study. All the dolomite is the typical buff, porous, cherty Niagaran type found along the west shore of Lake Michigan.

Devonian clastics from the basin of Lake Michigan form the great bulk of the pebbles of the axial portions of the Cary Lake Michigan moraines. The lithologies are identical to the Devonian of the Saginaw and Erie lobes. Brown-gray, well-indurated siltstone pebbles form 54.2% of the total, and gray-black, fissile New Albany Shale makes up 4.5%.

GREEN BAY LOBE: The pebble fraction of the till of the Green Bay lobe contains 12.0% Precambrian and 88.0% Paleozoic lithologies. Dolomite (82.4%) is present almost to the exclusion of other rock types (Fig. 2), as was also found by Thwaites (1943, Fig. 4). The provenance of the dolomite pebbles is probably threefold: (1) the Pairie du Chien Group (Ordovician), (2) the Galena Formation (Ordovician), and (3) the Niagara Dolomite (Silurian). All three formations parallel the western shore of Lake Michigan and extend in a broad belt from northern Michigan to southeastern Wisconsin. Since it is not possible to distinguish these individual sources among the pebbles, all the dolomites of the Green Bay lobe were grouped together.

Limestone pebbles are less abundant in the Green Bay lobe than in the Lake Michigan and constitute only 1.4% of the total. They are probably derived from the Ordovician limestone of Upper Michigan and eastern

Wisconsin, although the scarcity of limestone pebbles indicates that local limestone bedrock made few, if any, contributions to the drift. As in the lobes to the east the most abundant variety is the typical dense, brown to gray Middle Ordovician limestone (1.3%).

Grayish-brown, fine-grained Cambrian sandstone from the Dresbach Formation and brown, poorly indurated siltstone, perhaps from the Ordovician Maquoketa Formation, make up 1.2% of the total (Table 11).

DES MOINES LOBE: The pebble fraction of the till of the Des Moines lobe contains 34.6% Precambrian and 65.4% Paleozoic and Mesozoic lithologies. Limestone pebbles make up 21.4%, of which 19.2% are dense and buff gray (Fig. 2). The latter are probably derived from the local Mississippian, because the pebbles are similar to nearby outcrops and because no other extensive limestone areas were available to Des Moines ice. Other buff to brown porous limestone pebbles were found (2.2%), which could have been derived from Devonian, Mississippian, or Pennsylvanian bedrock in north-central Iowa.

Characteristic buff-white, cherty dolomite similar in every respect to the Silurian dolomites of the eastern lobes makes up 37.9% of the pebbles of the Des Moines lobe. Since the Ordovician dolomite of southeastern Minnesota made little contribution to the Des Moines lobe, the source of these pebbles must be the Ordovician and Silurian dolomites of the Red River valley in northwestern Minnesota and south-central Manitoba. Chert pebbles (0.5%) probably came from these northern dolomites and from the Mississippian limestones of north-central Iowa.

Clastic pebbles make up 5.5% of the total in the Des Moines lobe, of which 4.6% is derived from the Cretaceous shales of the Red River valley, southwestern Minnesota, and northeastern Iowa (Table 12). The shale pebbles are gray black, well indurated, and break with a characteristic conchoidal fracture. Ferrugenous concretions which were observed in association with the shales *in situ* were also found in the till and are therefore included with the Cretaceous constituents. The provenance of some of these, however, may be the Pennsylvanian of central Iowa. Gray-brown, poorly indurated siltstone pebbles (0.9%), because of their nonresistant nature, are believed to represent the local Pennsylvanian bedrock.

IOWAN DRIFT: The pebble fraction of the till of the Iowan drift contains 54.2% Precambrian and 45.8% Paleozoic and Mesozoic lithologies (Fig. 2). The limestone pebbles of the Iowan drift (17.7%) are lithologically similar to those of the Des Moines lobe, although there is less Mississippian, more Devonian, and little, if any, Pennsylvanian represented. The dense, buff-gray Mississippian limestones of the Des Moines lobe are found in the Iowan drift but to a lesser degree (7.0%). Likewise, buff, porous limestones are common to both drift sheets, but they are much more abundant in the Iowan (9.3%). These pebbles may be derived from the Devonian, since Iowan ice had access to a larger extent of Devonian bedrock. A gray, platy limestone (1.4%), peculiar to the Iowan, is probably also derived from the Devonian (Table 13).

There is significantly less dolomite in the Iowan (21.9%) than in the Des Moines lobe, but no lithologic differences are evident. The dolomite of the Iowan may have come from several sources: (1) the Red River valley, as postulated for the Des Moines lobe, (2) the the Ordovician of southeastern Minnesota, (3) the Silurian of northeastern Iowa. It has not been possible to distinguish pebbles from these various sources.

The clastic pebbles of the Iowan drift (3.9% are readily separable into two groups: (1) Cretaceous shales and ferruginous concretions and (2) Devonian and Carboniferous sandstones and siltstones. The former are identical to those in the Des Moines lobe and make up 2.7% of the total. The latter are not distinctive and give little evidence of their source. They are locally derived and form 1.1% of the total.

SUMMARY: The average of the pebble counts for all the lobes studied show that about 30% of the index grade size consists of far-traveled, Precambrian lithologies, and about 70% consists of Paleozoic and Mesozoic lithologies. The latter may be separated into three general classes: (1) lithologies distinctive of sources and restricted in their distribution which makes them useful index lithologies for certain regions, (2) lithologies distinctive of a particular source but common to all the lobes, (3) lithologies distinctive of neither a particular source nor a particular lobe.

On this basis the Paleozoic and Mesozoic lithologies of the six glacial lobes under consideration can be classified as follows:

Class (1): (a) the New Albany Shale, which occurs in the Erie, Saginaw, and Lake Michigan lobes and indicates an eastern source; (b) Cretaceous shale, which occurs in the Des Moines and Iowan drifts and indicates a

western source; (c) Ordovician limestones, which occur in the Erie, Saginaw, Lake Michigan, and Green Bay lobes and indicate an eastern source; (d) Devonian siltstone, which parallels the distribution of New Albany Shale and hence also indicates an eastern source.

Class (2): The dolomites, which are useful in differentiating lobes only in so far as they occur in differing amounts. In this respect the Green Bay (82.4%) and Des Moines (37.9%) lobes are distinctive; the others range between 21.9 and 26.2%.

Class (3): (a) Ordovician, Mississippian, and Pennsylvanian clastics; (b) all limestones except those representing the Ordovician Trenton and Black River.

Sand Lithology of the Lobes

GENERAL STATEMENT: The sand fraction (0.5–1.0 mm) was investigated at each locality where pebble samples were taken in order to corroborate the pebble-count data as well as to determine the effect of size reduction on till lithology. In every case the percentage of recognizable Paleozoic or Mesozoic lithologies (limestone, dolomite, sandstone, siltstone, shale) is smaller in the sand than in the pebble fraction; this is caused primarily by the concentration of quartz and feldspar in the finer grade sizes. The percentage of Precambrian lithologies (acidic, basic, and metamorphic fragments and mineral grains) is also much reduced in the sand fraction where many Precambrian pebbles are represented by grains of quartz and feldspar. Because the large amounts of quartz and feldspar in the sand fraction are derived from a variety of sources and because they are practically unknown in the pebble fraction, lithologic comparison between pebble and sand sizes can best be made by ratios.

Although the use of ratios in comparing sand and pebble counts will eliminate the effect of lithologies not common to both size grades, it will at the same time introduce variables which multiply the number of possible alternative explanations of the observed data. Thus where the ratio of two lithologies in the sand fraction is markedly different from the ratio of the same lithologies in the pebble fraction several alternative explanations are possible, depending upon the relative importance assigned to the following factors: (1) lithologic properties, especially those affecting durability in the sand and pebble sizes; (2) distance from source area; (3) rate of comminution; (4) dominant type of bedrock erosion by the ice, by plucking or abrasion.

Little attempt has been made to identify the source of sand grains, but the quartz-feldspar group was subdivided on the basis of rounding in order to indicate in a general way the relative importance of primary and secondary sources. These observations were made in lieu of more detailed investigation of the sand fraction, but it is hoped that they might indicate additional areas of profitable study.

TABLE 1.—COMPARISON OF RATIOS BETWEEN SAND AND PEBBLE FRACTIONS

Lobe	Precambrian/Paleozoic		Limestone/dolomite		Clastic/carbonate	
	Sand	Pebbles	Sand	Pebbles	Sand	Pebbles
Erie	0.12	0.32	2.08	1.91	0.78	0.14
Saginaw	0.32	0.64	1.16	0.85	0.44	0.35
Lake Michigan	0.05	0.12	0.31	0.10	1.21	2.05
Green Bay	0.33	0.14	0.03	0.02	0.14	0.01
Des Moines	0.19	0.53	0.88	0.57	1.40	0.09
Iowan	0.33	1.18	1.25	0.81	0.33	0.10

ERIE LOBE: The sand fraction of the Erie lobe consist of 66.8% Paleozoic lithologies but only 7.7% Precambrian (Fig. 2); this makes the ratio between the two much lower in the sand (.12) than in the pebble fraction (.32). Limestone and dolomite form 25.4 and 12.2% of the sand fraction, respectively, almost in the same relative proportions as was found in the pebble fraction (Table 1). Siltstone and shale fragments form 28.7% of the total and are much more abundant than in the pebble fraction. A quarter of the sand (25.5%) is quartz and feldspar, of which only 1.2% is rounded (Table 14).

SAGINAW LOBE: Paleozoic lithologies form 34.4% of the sand of the Saginaw lobe, about half as much as in the Erie lobe; Precambrian fragments make up 10.6% of the total, slightly more than in the Erie lobe (Fig. 2). The ratio of Precambrian to Paleozoic, however, is only half as large in the sand fraction as compared with the pebbles. The ratio of limestone (12.4%) to dolomite (10.7%) shows more variation from the pebble fraction (0.85) to the sand fraction (1.16) than does the Erie lobe. Clastic fragments are slightly more abundant, relative to carbonate fragments, in the sand. More than half of

the total (55.0%) consists of quartz and feldspar grains, of which 7.9% is rounded (Table 15)

LAKE MICHIGAN LOBE: Paleozoic lithologies make up 84.5% of the sand of the Lake Michigan lobe, more than in any other lobe studied (Fig. 2). Precambrian lithologies form only 4.2% of the total, which lowers the Precambrian-Paleozoic ratio from 0.12 in the pebble fraction to 0.05 in the sand. Limestone (8.8%) and dolomite (29.4%) are more common in the sand fraction than in the pebbles. Clastic fragments (46.2%) are less abundant in the sand (Table 16). The percentage of quartz and feldspar in the sand fraction of the Lake Michigan lobe is smaller than in any of the other lobes studied (11.4%).

GREEN BAY LOBE: The sand fraction of the Green Bay lobe contains 41.5% Paleozoic lithologies and 14.1% Precambrian lithologies, which, unlike any other lobe, cause the Precambrian-Paleozoic ratio to be larger in the sand fraction than in the pebbles. Limestone and dolomite make up 0.8 and 34.1%, respectively, of the total sand, and the ratio between the two is somewhat larger than in the pebble fraction. Compared to carbonate, clastic lithologies in the sand fraction (4.7%) are 14 times as abundant as similar lithologies in the pebble fraction (1.2%). Quartz and feldspar make up 44.4% of the sand, of which 17.9% is rounded (Table 16).

DES MOINES LOBE: In the Des Moines lobe Paleozoic lithologies constitute 36.4%, and Precambrian lithologies 7.1%, of the sand; the amount of Precambrian relative to Paleozoic is less in the sand than it is in the pebbles. Limestone and dolomite are present in almost equal amounts (7.2 and 7.7%, respectively), whereas in the pebble fraction there is almost twice as much dolomite. Siltstone and shale lithologies are much more common in the sand fraction (20.5%). Quartz and feldspar form 56.5% of the sand, of which 7.1% is rounded (Table 17).

IOWAN DRIFT: Paleozoic lithologies form only 12.4% of the sand of the Iowan drift, whereas Precambrian lithologies are even less abundant (4.2%). The Precambrian-Paleozoic ratio undergoes a very large reduction from the pebble to the sand size, 1.18 to 0.33. The Iowan drift has more Precambrian in the pebble fraction and less in the sand than any other lobe. This appears to be the case regardless of whether the sample was taken from outwash sand or ice-laid till. In common with every other lobe the Iowan has more limestone, relative to dolomite, in the sand fraction than in the pebbles. The clastic lithologies, relative to carbonate, are more abundant in the sand fraction. Quartz and feldspar make up 83.5% of the sand, far more than in any other lobe. Of this, 15.1% is rounded (Table 17).

LITHOLOGIC HOMOGENEITY WITHIN LOBES

GENERAL STATEMENT: In order to formulate lithologic indices capable of distinguishing till sheets it is necessary to investigate the variations within each lobe. Ideally for the purposes of differentiation a lobe should be lithologically homogeneous as well as lithologically distinctive. The present study indicates that neither of these conditions is ever perfectly obtained throughout a lobe and that the degree of lithologic homogeneity and distinctiveness varies not only from lobe to lobe but also among moraines of any given lobe. If, however, these variations can be interpreted in terms of local bedrock lithology or some other factor they may prove valuable rather than detrimental for purposes of correlation or differentiation. Therefore it appears unlikely that lithologic differentiation of glacial lobes can be accomplished in a purely mechanical manner, except on a very gross scale, for local variations must be carefully considered.

The Precambrian lithologies display less variation within lobes than do the Paleozoic and Mesozoic lithologies because they are farther from the source and, as is shown by the general similarity between lobes, because local variations in the source areas are not so strongly marked. The lithologic variations in the sand fraction parallel those of the pebble fraction, but they are usually less pronounced in the sand. This may indicate that the sand fraction is not so greatly overloaded in the lee of an outcrop as is the pebble fraction, or perhaps it merely indicates that local lithologic variations in the sand are dampened by the concentration of quartz and feldspar in the finer sizes. The widely spaced samples of this study give only a rough approximation of the areal lithologic variations, and it is quite possible that each end and ground moraine unit may exhibit areal variations different from those of neighboring moraines.

The lithologic homogeneity within each lobe can be seen in the accompaning tables and need not be discussed at length. However, certain anomalies in both the total and Precambrian counts may be pointed out.

TOTAL COUNTS: The variations in the total counts within the lobes are due largely to local variations in the source area, involving changes

in lithology, topographic relief, and extent of older drift cover. Pronounced local variations caused by changes in bedrock lithology are numerous, but they cannot always be traced to a definite area of outcrop. In the following cases the relationship is apparent: (1) the increase in dolomite in the southern limbs of the three outer moraines of the Erie lobe (samples 29, 30, 31) is caused by near-by exposures of Silurian dolomite along the Wabash valley; (2) the high per cent of clastics in sample 33 (51%) from the Maxinkuckee moraine of the Saginaw lobe is the result of proximity to Devonian siltstone and shale; (3) Silurian dolomite at the south end of Lake Michigan caused marked variations among the samples from the Lake Michigan lobe; (4) Silurian dolomite exposed along the Wapsipinicon River in eastern Iowa is responsible for the increase in dolomite in the Iowan drift noted in sample 12; (5) Devonian limestone in north-central Iowa accounts for the high limestone percentages in the northern samples from the Des Moines lobe.

In other cases local variations cannot be traced to definite bedrock sources because no near-by exposures are now evident, or because the bedrock yields pebbles which are not distinctive and which may be ascribed to a number of local formations. Most of the variations in the Saginaw lobe are of this type. Here, however, the variable nature of the underlying Carboniferous sediments no doubt played a large role in causing variability in the lithology of the till. The high limestone percentages in the Mankato Port Huron moraine (samples 51, 52) may be the result of the till not being as fully contaminated with Carboniferous clastics as it is farther south, or they may be due to changes in regimen or direction of movement of the ice sheet, or both, during the Mankato advance, which subjected the limestone area in the Georgian Bay region to more vigorous glacial action.

PRECAMBRIAN COUNTS: With the exception of the Lake Michigan and Green Bay lobes, the Precambrian pebble counts display only slight variations within lobes. The variations which do exist have little apparent system or trend, but in some cases they can be referred to local sources. In other cases, especially in the Lake Michigan lobe, no explanation for the lack of consistency is readily apparent.

No notable variations were found within the Erie or Saginaw lobes. The Lake Michigan lobe exhibits wide variations in Precambrian lithology which have no apparent relationship to specific moraines. All three major Precambrian types show large anomalies, for which the data at hand cannot account. The Green Bay lobe is homogeneous, except for sample 70 which is high in basalt and sample 72 which is high in quartzite. There is no known source for the basalt in sample 70, but it must be local because the other samples in that area do not carry large amounts of basalt. Sample 72 is in the lee of quartzite outcrops near Waterloo, Wisconsin, which overloaded the till with quartzite pebbles. The Precambrian lithologies of the Des Moines lobe are remarkably constant, although samples 90 and 91 have a slight excess of granite. This is the result of their proximity to granite exposures along the valley of the Minnesota River. The Iowan samples are less consistent than those from the Des Moines lobe, but no large anomalies are evident.

Interpretation of Drift Lithology

GENERAL STATEMENT: A previous study of pebble counts from the Marseilles moraines in northeastern Illinois (Anderson, 1955) indicated that the following factors influence the distribution of any given lithologic type in the pebble fraction of glacial till: (1) provenance—the extent and proximity of the source area, its topographic position, and the amount of cover by older drift; (2) lithologic properties—jointing, weathering products, resistance to glacial erosion; (3) glacial processes—the regimen of the ice sheet and the vertical and horizontal distribution of the rock types within the ice; and (4) dilution—the addition of other lithologic types. These factors also determine the lithology of the sand fraction, but probably the relative importance of each varies from one size grade to the next. Consideration of sand grain and Precambrian pebble counts permits a clearer interpretation of drift lithology and thus a more accurate determination of lithologic differences among the various till sheets.

ERIE LOBE: Ordovician limestones are more abundant in the Erie than any of the other eastern lobes, even though the source area in this case is the most distant. Several factors appear to be responsible for this situation: (1) the source area of the limestone surrounding Lake Ontario is more extensive than that available to Saginaw, Lake Michigan, or Green Bay ice; (2) the rock is resistant enough to have survived the trip from Lake Ontario to northeastern Indiana in great numbers; (3) other resistant lithologies were not added en

route to reduce appreciably the percentage of Ordovician limestone. The shale from the Lake Erie basin was unable to withstand the rigors of glacial transport, and most of it was removed from the pebble fraction by the time the ice had advanced to northeastern Indiana. At the same time small amounts of Silurian dolomite were added in northwestern Ohio and along the Wabash River in northeastern Indiana. Subdued bedrock relief and a protective cover of older drift may have prevented the Silurian from making a larger contribution in this area.

The sand fraction substantiates the fact that the Devonian shales underwent extensive comminution during transport, for, relative to carbonate, they are about 5½ times as abundant as in the pebble fraction. The small percentage of rounded quartz and feldspar grains (1.2) suggests that sedimentary rocks were not important sources for quartz and feldspar. Most of these grains were derived from the disintegration of primary rocks.

Precambrian pebbles, though far removed from their source, make up nearly a quarter of the total in the Erie lobe. Of this, 54.6% is diorite, greenstone, basalt, and gabbro, which forms only a small fraction of the bedrock of the source area. Evidently they are common in the till 450 miles from the source because of the selective elimination of the weaker granites, gniesses, and schists. Gabbro and other coarse-grained igneous rock are more common in the Erie than in any other lobe (8.4%), which may be a reflection of the comparatively large areas of basic intrusives in southern Quebec. The 20.2% graywacke, slate, schist, and quartzite is probably representative of the Grenville Series of southwestern Quebec and southeastern Ontario. Acidic rocks form only 20.2% of the Precambrian, a figure that is low compared to their abundance in the source area. In the Erie till, as in all the other tills studied except the Iowan, quartz forms a minor part of the Precambrian fraction. The predominantly angular pebbles indicate that primary igneous rocks were the most important sources for the quartz. Rounded quartz pebbles are rare but ubiquitous in their distribution throughout the lobes studied. M. M. Leighton suggested (Personal communication) that the source for the rounded quartz pebbles may be found in remnants of high-level pre-Pleistocene gravels which have been reported throughout most of the Central Lowland. (*See* Potter, 1955, p. 1–3.)

SAGINAW LOBE: The percentage of limestone in the Saginaw lobe (19.6) is much lower than in the Erie lobe, although the lithologies are very similar. Two factors appear responsible for this reduction in the percentage of limestone: (1) the source area for Ordovician limestone pebbles is of smaller extent in the Georgian Bay region, compared with the region adjoining Lake Ontario; (2) the addition of large amounts of other lithologies, notably Silurian dolomite and Carboniferous clastics, tends to reduce the percentage of limestone.

The clastic lithologies of the Saginaw lobe, chiefly Mississippian and Pennsylvanian siltstone and shale, are more abundant than the Devonian shale and siltstone in the Erie lobe because they are in a more favorable topographic position and are close to the site of deposition. They are not resistant, however, and appear to have influenced till composition for only a few miles downstream from an eroded outcrop. The dolomite pebbles of the Saginaw lobe were probably derived entirely from Silurian exposures in the Georgian Bay region.

The higher percentage of Precambrian pebbles in the Saginaw lobe compared with the Erie is apparently the result of a closer source area. The Huronian rocks of the north shore of Lake Huron are numerous in the till and are probably the cause of the decrease in the percentage of granite and basic rocks. Higher percentages of greenstone and quartzite reflect the proximity of Huronian sources, but the rare pebbles of tillite(?) and jasper conglomerate point more conclusively to this source. Slosson (1933) demonstrates that the source of the jasper conglomerate is localized in the Bruce Mines area, Ontario.

There is a smaller percentage of clastic fragments in the sand fraction than in the pebbles, which indicates either that the local siltstones and shales did not undergo extensive comminution during glacial transport or that the sand is overloaded with quartz and feldspar from the local sandstones. The percentage of quartz and feldspar is more than twice as large (55.0%) as in the Erie lobe, and the percentage of rounded quartz and feldspar grains (7.9) is almost 7 times as large, which indicates that sedimentary sources were more important in the Saginaw lobe. Certainly much of the increased percentage of quartz and feldspar, especially the angular grains, must be attributed to disintegration of larger amounts of Precambrian rocks. The larger limestone-dolomite ratio in the sand fraction indicates that the limestone is farther from its source than is the dolomite, or that, if differences in

the distance of transport are insignificant, the dolomite pebbles are more resistant than the limestone and thus are not reduced in size as readily.

LAKE MICHIGAN LOBE: The Cary moraines of the Lake Michigan lobe are in a favorable location relative to the Devonian clastics of the Lake Michigan basin. The great distance through which Lake Michigan ice overrode the Devonian, the easily eroded character of these siltstones and shales, and the absence of other bedrock lithologies together make possible the preponderance of Devonian clastics in the Cary moraines of the Lake Michigan lobe. Notwithstanding the high percentage of clastic lithologies, dolomite is more common than in either the Saginaw or Erie lobes. It appears for the most part to be derived locally from the Niagaran at the south end of the lake. Although there is a potential source in the Mackinac Straits region, this probably contributed only a small portion of the dolomite pebbles. The distinctive Ordovician limestones of the Straits area form only 1.7% of the total at the south end of the lake, and the writer assumes that a comparable amount of dolomite was also derived from that source.

The great influx of Devonian siltstone and shale is undoubtedly the main cause for the lower percentage of Precambrian pebbles in the Lake Michigan lobe. The major contrast in Precambrian lithologies between the Lake Michigan and the Saginaw and Erie lobes are caused by the scarcity of Huronian and Grenville metasediments in the former, coupled with a substantial increase in the percentage of granite and basalt. Therefore the ice of the Lake Michigan lobe appears to have by-passed the main Huronian area north of Lake Huron and may have acquired more granite along a route lying farther northwest. The increase in the percentage of basalt might be due to differential resistance of the Precambrian lithologies to glacial abrasion, since basalt is extremely resistant. Another alternative is an actual increase in the amount of basalt acquired in the source area, although there is no evidence that basalt is more prevalent along the path of Lake Michigan advance than it is in the source areas of the lobes to the east.

The fact that the ratio of clastic to carbonate fragments is larger in the pebble fraction than in the sand, as well as the fact that there is a smaller percentage of clastics in the sand fraction, suggests that they have been derived from local sources contributing larger particles, which as yet have not traveled far enough to be fully represented in the sand sizes. Differences in lithologic properties are probably not great enough to account for the fact that the limestone-dolomite ratio in the sand fraction is 3 times that of the pebble fraction. It is more likely that the dolomite pebbles were added from local sources and, like the Devonian clastics, have not traveled far enough to be fully represented in the sand fraction, whereas the limestone has suffered considerable size reduction during transport from the north end of Lake Michigan. Several factors account for the high percentage of Paleozoic lithologies in the sand fraction: (1) overloading with Devonian clastics, (2) absence of Precambrian sources within at least 400 miles, (3) limited sandstone sources curtailing the supply of quartz and feldspar which elsewhere occur in great abundance. More than a quarter of the quartz and feldspar is rounded, which indicates an important sedimentary source, probably Keweenawan and Cambrian sandstones.

GREEN BAY LOBE: The combination of near-by source and resistant lithology appears to be responsible for the concentration of dolomite pebbles in the Green Bay lobe and the concomitant reduction of the percentage of Precambrian lithologies below that of any other lobe, even though Precambrian sources are near by. Bedrock occurrences of limestone and clastics along the line of ice movement are very restricted and, except for the typical Ordovician limestones, are neither resistant nor distinctive. Hence they are of minor importance.

The distribution of the Precambrian lithologies is more distinctive in the Green Bay lobe than in any other of this study. The contrast with the neighboring Lake Michigan lobe is striking, since the line of movement of Green Bay ice was far enough west to permit a sizable load of Lake Superior lithologies to be acquired. Granite, schist, and basic rocks, especially the dioritic types, are replaced by increased amounts of rhyolite, quartzite, gneiss, jasper, iron formation, and Keweenawan and Cambrian red clastics which were derived primarily from sources in Wisconsin and Upper Michigan. The comparatively high percentage of quartzite does not occur uniformly throughout the lobe but is concentrated in the southern portions of the lobe, as has been pointed out by Alden (1918, p. 253). Thus the high amount of quartzite (12.5%) is the result of the inclusion of three samples taken in the lee of quartzite ledges in southwestern Dodge County

near Waterloo, Wisconsin (Buell, 1895). North of these local sources the amount of quartzite averages about 2% of the Precambrian, or nearly the same as in the Lake Michigan lobe.

The sand fraction of the till of the Green Bay lobe lends support to the observation that the local limestone and clastic lithologies do not readily form pebbles but are instead found in the finer size grades. About 40% of the quartz and feldspar (or about 18% of the sand) is rounded and may have a sedimentary origin, probably the Cambrian sandstones. Shale fragments, probably from the Ordovician Maquoketa Formation, are much more abundant in the sand fraction. The ratio of limestone to dolomite shows a slightly greater relative amount of limestone in the sand. This may be due to either the weakness of the limestone or the proximity of the dolomite contributing larger amounts of pebble-sized fragments.

DES MOINES LOBE: The large amount of dolomite found in the Des Moines lobe was evidently brought southward from Ordovician and Silurian sources in southern Manitoba, since no other source is readily apparent. The limestone is locally derived from Mississippian exposures. The relatively low percentage of limestone and its uniform distribution seem to indicate a protective cover of older drift. Cretaceous shales do not form a conspicuous part of the pebble fraction (4.2%), because of their nonresistant character and the distance from bedrock sources. Likewise, little of the local Pennsylvanian is represented in the till, although this may be due in part to a protective cover of older drift. Zumberge and Wright (1956, p. 68–69) found 29% shale in the 2- to 8-mm size grade of the "youngest" (Mankato) till in southern Minnesota.

The Precambrian pebbles of the Des Moines lobe show striking contrasts with the eastern lobes: (1) restricted amounts of quartzite and diorite, (2) significantly greater amounts of graywacke and granite, (3) an almost total absence of Lake Superior lithologies. Considering its proximity and high topographic position, the Sioux Quartzite is present in surprisingly small quantity. However, increased numbers of granite and schist pebbles were found whose source is probably the Archean granites and schists exposed along the Minnesota River in southwestern Minnesota with possibly minor additions from areas of Algoman granite and schist in the western part of the state and from more distant Shield areas. If this is the case it is difficult to explain why the granite in the Minnesota valley was extensively eroded by the ice, whereas the Sioux Quartzite on near-by topographic highs was not. M. M. Leighton (Personal communication) suggested that the Sioux Quartzite lay in the zone of lateral ice movement, and as a result it is not abundant in the axial portions of the lobe. The comparatively high percentages of graywacke and slate were derived from the Animike and Knife Lake groups in western Minnesota.

More than half of the sand fraction is quartz and feldspar, but only 7.1% is rounded, which indicates that sedimentary sources were subordinate to igneous and metamorphic sources. Fragments of Cretaceous shale are abundant in the sand fraction because they are sufficiently far removed from the source area to have undergone considerable comminution. Shale, quartz, and feldspar occur in such abundance that limestone and dolomite are relatively inconspicuous in the sand fraction.

IOWAN DRIFT: Ice of the Iowan advance passed over more Devonian and Mississippian limestone than did the Des Moines ice, and, although the till of the latter has a higher percentage of limestone, the Iowan drift has more limestone relative to the total amount of Paleozoic and Mesozoic pebbles. Since there is evidence that Iowan ice advanced over a thick cover of weathered Kansan till in northeastern Iowa, limestone probably was not incorporated into the till in amounts commensurate with its occurrence in the bedrock. Dolomite is less abundant compared to the Des Moines lobe both in relation to the Paleozoic as well as to the total, probably because of the influx of limestone and Precambrian pebbles. The source of the dolomite is not definitely known, since the Ordovician and Silurian of southern Manitoba, the Ordovician of southern Minnesota, and the Silurian of northeastern Iowa were all possible contributors. Because they are farther from the source, Cretaceous shales are not so common as in the Des Moines lobe, but siltstone and shale from the local Devonian and Carboniferous are somewhat more plentiful.

The pebble counts from the Iowan drift have affinities to the lithologies of Lake Superior as well as to those of western Minnesota. There are two possible explanations for this anomaly: (1) the Iowan drift of eastern Iowa is the product of the Red-Iowa lobe (Horberg and Anderson, 1956, p. 110) which during Iowan time may have deployed widely enough to incorporate Precambrian lithologies from central Minnesota as well as Cretaceous shales

from western Minnesota and North Dakota and South Dakota; (2) Iowan drift was deposited by ice moving from the northwest, perhaps a source west of the traditional Keewatin center (Horberg and Anderson, 1956, p. 110), but reworked an earlier drift sheet containing Lake Superior lithologies. Higher percentages of granite, graywacke, and schist, as well as Cretaceous shales, indicate a source to the northwest similar to that of the Des Moines lobe, whereas comparatively high percentages of rhyolite and greenstone (presumably the Ely Greenstone of north-central Minnesota) and occurrences of jasper and red clastics point to a Lake Superior source. The problem must remain unanswered until more data have been collected, especially from the older drifts. The Iowan drift contains almost twice as much greenstone as any other lobe, and therefore Iowan ice must have had either a unique line of advance or a unique source of greenstone, or both. The greenstone may have come directly from the Ely Greenstone and other Precambrian sources or may have been reworked from a soil profile on Kansan till in which they are found as residuals. The latter process may have operated in bringing about the concentration of quartz in the Iowan drift.

The extremely high percentage of quartz and feldspar in the sand fraction further accentuates the problem of the source of the Iowan. The quartz could be derived (1) from a reworked soil profile, (2) from the arkosic Keweenawan and Cambrian sandstones of eastern and northern Minnesota, or (3) from pre-Pleistocene gravels. The small number of rounded grains suggests that the former possibility is more reasonable. Little information can be gained from the other sand lithologies because of their relative scarcity.

Pebble Counts from Kansan Till

Counts were made of pebbles collected by Leland Horberg and the writer from Kansan till in western Illinois and eastern Iowa in order to determine, if possible, the source of the till solely on lithologic evidence. It is generally accepted that this till was deposited by ice moving from the Keewatin center. (*See* Horberg, 1950, p. 100.) The pebble counts are listed and summarized in Tables 18 and 19.

Dolomite forms 18.7% of the total and is similar to the Ordovician and Silurian types found in all the Wisconsin lobes. The limestones (14.5%) are buff to brown; the majority are fine-grained and dense. They are similar in all respects to the Devonian and Mississippian limestones found in the Iowan drift and Des Moines lobe. Clastic lithologies, poorly indurated siltstones, make up 10.3% of the total and are probably derived from the local Pennsylvanian bedrock. No characteristic Creataceous shales were found in the Kansan samples, although Udden (1899) reported their presence in eastern Iowa. The pebbles are 55% Paleozoic.

Granite and various types of basic rocks form 49.2% of the Precambrian pebbles. The granites (19.0%) are of the typical Laurentian variety which occurs throughout the drift of the Central Lowland. Of the basic rocks, the fine-grained types are most abundant (16.0%), whereas coarse-grained types (5.5%), medium-basic types (8.7%), and greenstone (6.8%) are less common. The metamorphic rocks, quartzite (5.5%), schist (9.2%), and graywacke (17.5%), are lithologically similar to those found in the Wisconsin till. Quartz (6.0%), rhyolite, and thyolite porphyry (2.3%), jasper (0.8%), and red clastics (0.2%) also occur.

Although certain characteristics of the Kansan counts are common to several of the Wisconsin lobes, others show strong affinities toward a western source. The limestones offer positive evidence that the Kansan ice which deposited this till moved across the Devonian and Mississippian limestones of eastern Iowa (Horberg, 1956, p. 20). The percentage of graywacke in the Precambrian counts is much greater than that found in any eastern lobe, but it compares favorably with the Iowan and Des Moines lobes. The presence of Lake Superior lithologies, jasper, rhyolite, and red clastics links the Kansan more specifically to the Iowan, for such lithologies are practically unknown in the Des Moines lobe. Whether this indicates a similarity of source for the Kansan and Iowan, or indicates that the Lake Superior lithologies of the latter were reworked from the former, cannot be shown from these data.

Kansan till of eastern Iowa and western Illinois and Iowan till of eastern Iowa contain notable amounts of quartz pebbles. Since these are the only tills sampled which are known to have overridden extensive areas of older drift on which interglacial soils have developed—*i.e.*, the Kansan overrode an Aftonian soil profile developed on Nebraskan drift, and the Iowan overrode a complex Yarmouth-Illinoian-Sangamon soil profile developed on Kansan drift—it seems logical to conclude that these larger amounts of quartz pebbles have been secondarily derived as residuals from the underlying soil profile. Several lines of evidence sup-

port this conclusion: (1) the pebbles are not rounded or patinated, characteristics to be expected if the source were a pre-Pleistocene gravel; (2) the relatively high percentages of quartz are found throughout the area sampled, which indicates a widespread rather than localized source; (3) the association with larger amounts of resistant Precambrian lithologies coupled with smaller percentages of carbonates, compared with the Des Moines lobe which passed over much the same bedrock. Pebble counts cited by Kay and Apfel (1929) indicate the extent to which quartz is concentrated in deeply weathered soil profiles: percentage of quartz in unleached Nebraskan till in Iowa—3.3; in Nebraskan gumbotil—36.75; in unleached Kansan till—2.0; in Kansan gumbotil—48.5. Hence the larger amounts of quartz pebbles in Iowan and Kansan till do not necessarily indicate a similarity of source or direction of ice movement.

Conclusions

(1) Significant lithologic differences exist among the major Wisconsin glacial lobes of the Central Lowland.

(2) The far-traveled, Precambrian lithologies are valuable for differentiating lobes because they are usually uniformly distributed throughout a given lobe and because recognizable quantitative, and in some places qualitative, differences occur between lobes.

(3) Local Paleozoic and Mesozoic constituents display greater variations between lobes than do the far-traveled pebbles, but they also show greater variations within a single lobe. In some cases this detracts from their usefulness as criteria in distinguishing drift sheets, especially where lithologies appear which cannot be linked to a specific bedrock source.

(4) Counts of the sand fraction are useful in corroborating the pebble data, but, taken alone, they indicate little about the source of the till, since the lithologic properties of the grains cannot be accurately determined. They appear to be more consistent throughout a lobe than do the pebble counts and may in some instances be valuable in differentiating tills where the question of source is not involved.

(5) The relative amounts of a given lithology in the sand- and pebble-grade sizes are dependent upon the following factors: (a) lithologic properties, especially those affecting durability in the sand and pebble sizes; (b) distance from source; (c) rate of comminution; (d) dominant type of glacial bedrock erosion.

(6) The consistency of the total pebble counts without a great predominance of one lithologic type in the Erie and Des Moines lobes suggests that local bedrock did not make significant contributions to the till. This would indicate the presence of a cover of older drift, if the bedrock in question was within the zone of glacial erosion.

(7) Iowan ice moved across the weathered surface of the underlying Kansan till and incorporated large amounts of Precambrian residuals from the soil profile into the Iowan drift. Whether the Lake Superior lithologic suite of the Iowan drift was secondarily derived from the Kansan or brought directly by Iowan ice moving southward along the axis of the Red-Iowa lobe cannot be determined from the data at hand.

(8) The Kansan till of western Illinois and eastern Iowa has strong affinities to both the Des Moines and Iowan drifts, but the similarities to the latter are the most pronounced.

(9) The relatively large amounts of quartz in both the Kansan and Iowan samples were derived principally as residuals from the underlying soil profiles.

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Manuscript Received by the Secretary of the Society, March 4, 1957

Table 2.—Counts of Precambrian Pebbles from the Erie Lobe

	Sample	22	23	24	25	26	27	28	29	30	31	32	21	Total	Average
Granite	Gneissic	12	11	5	20	24	12	10	8	4	10	8	8	132	11.0
	Laurentian	7	4	6	6	5	13	15	4	10	13	5	12	100	8.3
	Total	19	15	11	26	29	25	25	12	14	23	13	20	232	19.3
Basic	Coarse-grained	4	3	15	16	6	12	11	5	2	3	5	19	101	8.4
	Fine-grained	23	22	25	24	21	19	32	23	18	18	20	21	266	22.2
	Fine-grained, banded	1	2	2	0	0	4	0	1	0	5	6	0	21	1.8
	Total	28	27	42	40	27	35	43	29	20	26	31	40	388	32.4
Medium basic		21	27	15	11	15	11	8	21	32	17	22	14	214	17.8
Metamorphic	Gneiss	4	5	6	8	15	1	2	2	4	2	5	5	59	4.9
	Graywacke	10	7	4	3	1	2	6	12	2	9	8	7	71	5.9
	Hornblende schist	2	2	2	5	8	5	2	1	0	3	3	0	33	2.8
	Quartzite	10	9	12	3	0	13	7	13	17	12	10	14	120	10.0
	Slate	1	2	1	3	4	2	4	1	0	0	0	0	18	1.5
	Total	27	25	25	22	28	23	21	29	23	26	26	26	301	25.1
Greenstone		4	6	7	0	0	4	2	5	9	8	8	0	53	4.4
Vein quartz		1	0	0	1	0	2	1	4	2	0	0	0	11	0.9
Rhyolite		0	0	0	0	1	0	0	0	0	0	0	0	1	0.1
Total		100	100	100	100	100	100	100	100	100	100	100	100	1200	100.0

TABLE 3.—COUNTS OF PRECAMBRIAN PEBBLES FROM THE SAGINAW LOBE

	Sample	17	18	19	20	33	36	37	38	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	Total	Average
Granite	Gneissic	1	3	0	0	7	2	1	3	1	0	2	1	0	5	0	2	0	4	7	5	3	0	0	6	3	0	3	2	5	0	1	67	2.2
	Laurentian	17	24	36	17	19	8	10	10	23	13	12	12	21	14	13	11	18	23	8	8	15	9	16	16	24	12	13	11	16	9	6	464	14.9
	Total	18	27	36	17	26	10	11	13	24	13	14	13	21	19	13	13	18	27	15	13	18	9	16	22	27	12	16	13	21	9	7	531	17.1
Basic	Coarse-grained	6	7	1	11	4	7	2	3	1	0	0	7	3	3	4	13	7	1	9	7	6	2	2	5	5	4	5	5	1	5	3	139	4.5
	Fine-grained	11	10	9	17	19	19	20	17	8	10	3	21	15	19	15	14	15	17	17	31	25	26	21	8	16	21	20	28	10	23	15	520	16.8
	Fine-grained, banded	4	0	2	1	1	1	7	3	2	6	4	3	2	4	2	1	4	0	4	3	4	0	0	2	1	2	0	2	7	2	4	78	2.5
	Total	21	17	12	29	24	27	29	23	11	16	7	31	20	26	21	28	26	18	30	41	35	28	23	15	22	27	25	35	18	30	22	737	23.8
Medium basic		15	32	22	18	26	18	23	27	24	22	46	24	17	25	23	23	21	25	20	19	23	27	27	30	20	21	17	16	16	16	17	700	22.6
Metamorphic	Gneiss	5	8	0	5	6	4	2	2	2	2	4	1	3	0	5	3	7	3	3	2	2	3	8	1	3	0	2	6	4	2	4	102	3.3
	Graywacke	10	1	3	5	5	12	10	6	6	8	7	8	7	8	8	4	5	3	4	5	5	5	3	10	1	6	5	4	8	11	11	194	6.3
	Hornblende schist	0	2	0	5	4	4	2	1	3	2	0	5	2	3	1	3	3	3	3	3	2	3	0	3	6	3	2	4	3	0	0	75	2.4
	Green schist	0	3	0	1	0	2	0	1	1	0	1	0	0	0	1	0	0	0	1	0	0	1	0	1	0	0	0	0	0	1	0	14	0.5
	Quartzite	13	6	8	11	4	17	16	19	16	17	9	11	23	14	22	21	12	19	13	10	13	13	13	11	17	22	22	14	17	19	19	461	14.8
	Slate	0	0	0	1	0	0	0	0	0	3	0	1	1	1	0	0	1	0	0	0	1	0	0	0	0	1	0	0	0	0	0	10	0.3
	Total	28	20	11	28	19	39	30	29	28	32	21	26	36	26	37	31	28	28	24	20	22	26	24	26	27	32	31	28	32	33	34	856	27.6
Greenstone		11	2	17	7	4	3	7	4	10	15	11	5	3	4	2	1	6	1	9	6	2	8	6	6	4	8	8	6	6	10	12	204	6.6
Vein quartz		1	0	1	0	1	1	0	2	2	1	1	1	1	0	2	3	1	1	2	1	0	0	2	1	0	0	1	0	5	0	5	36	1.2
Rhyolite		3	2	1	0	0	1	0	0	1	0	0	0	1	0	0	0	0	0	0	0	0	0	1	0	0	0	0	0	0	1	0	11	0.4
Keweenawan clastics		0	0	0	1	0	0	0	0	0	0	0	0	0	0	2	0	0	0	0	0	0	1	0	0	0	0	0	0	0	1	0	5	0.2
Jasper		1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	Tr
Jasper conglomerate		0	0	0	0	0	1	0	1	0	1	0	0	1	0	0	1	0	0	0	0	0	1	1	0	0	0	2	2	1	0	1	13	0.4
Tillite (?)		2	0	0	0	0	0	0	1	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0	2	6	0.2
Total		100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	3100	100.0

TABLE 4.—COUNTS OF PRECAMBRIAN PEBBLES FROM THE LAKE MICHIGAN LOBE

Sample		1	2	4	5	6	7	8	9	10	11	15	16	34	35	Total	Average
Granite, Laurentian		33	19	35	27	25	20	25	15	27	29	23	34	19	20	351	25.1
Basic	Coarse-grained	1	6	7	1	3	1	0	3	0	1	3	4	9	4	43	3.1
	Fine-grained	38	36	19	35	49	49	48	39	21	20	28	5	34	26	447	31.9
	Fine-grained, banded	0	0	0	0	0	0	0	0	0	0	0	0	1	0	1	0.1
	Total	39	42	26	36	52	50	48	42	21	21	31	9	44	30	491	35.1
Medium basic		11	14	10	16	10	9	9	12	15	18	37	18	9	23	211	15.1
Metamorphic	Gneiss	0	0	1	2	1	0	2	5	6	1	1	2	2	3	26	1.9
	Graywacke	0	5	8	7	1	1	3	4	9	10	0	14	9	9	80	5.7
	Hornblende schist	11	10	8	5	6	14	3	11	6	4	2	8	6	8	102	7.3
	Green schist	0	0	2	0	0	0	0	0	0	0	0	4	3	1	10	0.7
	Quartzite	2	1	2	1	2	3	0	6	4	2	3	5	3	1	35	2.5
	Total	13	16	21	15	10	18	8	26	25	17	6	33	23	22	253	18.0
Greenstone		3	3	3	2	0	2	4	1	2	1	2	6	5	2	36	2.6
Vein quartz		0	3	2	1	1	0	0	3	0	6	0	0	0	1	17	1.2
Rhyolite		0	0	0	0	2	1	0	0	4	3	1	0	0	0	11	0.8
Rhyolite porphyry		0	0	0	0	0	0	0	0	0	1	0	0	0	0	1	0.1
Keweenawan clastic		1	3	2	2	0	0	5	1	4	4	0	0	0	2	24	1.7
Jasper		0	0	1	1	0	0	1	0	2	0	0	0	0	0	5	0.4
Total		100	100	100	100	100	100	100	100	100	100	100	100	100	100	1400	100.0

TABLE 5.—COUNTS OF PRECAMBRIAN PEBBLES FROM THE GREEN BAY LOBE

	Sample	63	64	64	66	68	69	70	71	72	73	74	Total	Average
Granite	Gneissic	3	0	1	2	0	0	2	2	0	3	6	19	1.7
	Laurentian	16	14	9	9	12	10	4	12	8	14	16	124	11.3
	Total	19	14	10	11	12	10	6	14	8	17	22	143	13.0
Basic	Coarse-grained	4	8	3	5	4	13	18	6	2	1	7	71	6.5
	Fine-grained	19	30	23	27	34	24	51	20	1	32	19	280	25.4
	Fine-grained, banded	1	0	3	1	0	0	0	3	1	0	0	9	0.8
	Total	24	38	29	33	38	37	69	29	4	33	26	360	32.7
Medium basic		9	7	25	13	10	7	2	12	5	11	8	109	9.9
Metamorphic	Gneiss	3	3	9	5	8	2	2	5	1	3	8	49	4.5
	Graywacke	8	8	9	7	0	3	4	9	5	13	9	75	6.8
	Hornblende schist	2	4	4	1	0	0	3	1	0	2	3	20	1.8
	Green schist	0	0	0	1	0	0	0	0	1	0	0	2	0.2
	Quartzite	7	8	10	16	0	4	2	15	67	3	5	137	12.5
	Total	20	23	32	30	8	9	11	30	74	21	25	283	25.8
Greenstone		4	2	0	1	4	8	1	0	2	0	1	23	2.1
Vein quartz		0	3	0	0	0	0	0	0	2	1	6	12	1.1
Rhyolite		3	1	2	1	0	7	1	4	1	2	2	24	2.2
Rhyolite porphyry		3	0	0	2	8	5	2	1	3	4	1	29	2.6
Keweenawan clastics		10	10	2	8	14	17	7	10	1	6	6	91	8.3
Iron formation		1	0	0	0	6	0	0	0	0	0	1	8	0.7
Jasper		7	2	0	1	0	0	1	0	0	5	2	18	1.6
Total		100	100	100	100	100	100	100	100	100	100	100	1100	100.0

TABLE 6.—COUNTS OF PRECAMBRIAN PEBBLES FROM THE DES MOINES LOBE

	Sample	77	78	79	80	81	82	83	84	85	86	87	88	89	90	91	92	93	Total	Average
	Granite, Laurentian	31	30	26	20	24	28	37	27	27	33	32	26	25	39	46	37	26	514	30.3
Basic	Coarse-grained	3	6	5	4	2	7	3	4	5	7	7	11	7	7	7	12	8	105	6.2
Basic	Fine-grained	28	27	21	19	29	17	15	21	17	8	24	23	21	12	11	14	22	329	19.3
Basic	Fine-grained, banded	1	0	1	0	0	2	2	0	4	0	1	0	0	0	0	0	0	11	0.6
Basic	Total	32	33	27	23	31	26	20	25	26	15	32	34	28	19	18	26	30	445	26.1
Medium basic		9	4	14	14	3	16	6	15	4	15	7	8	14	13	4	3	2	151	8.9
Metamorphic	Gneiss	1	0	3	2	2	2	0	0	2	0	1	1	0	0	0	2	2	18	1.1
Metamorphic	Graywacke	8	18	13	17	10	8	16	11	25	17	15	11	9	10	18	23	23	252	14.8
Metamorphic	Hornblende schist	6	5	7	11	7	9	8	8	10	5	6	10	10	6	5	4	8	125	7.4
Metamorphic	Green schist	1	0	0	0	0	1	0	0	0	0	1	1	0	0	0	0	0	4	0.2
Metamorphic	Quartzite	7	3	3	0	9	3	1	6	1	3	0	6	3	4	4	3	4	60	3.5
Metamorphic	Slate	0	2	2	0	0	0	0	0	0	0	0	0	0	0	0	0	0	4	0.2
Metamorphic	Total	23	28	28	30	28	23	25	25	38	25	23	29	22	20	27	32	37	463	27.2
Greenstone		5	3	5	10	12	4	9	6	2	12	6	3	6	7	4	2	3	99	5.8
Vein quartz		0	0	0	0	0	1	1	1	3	0	0	0	4	0	0	0	0	10	0.6
Rhyolite		0	1	0	1	0	1	2	1	0	0	0	0	1	0	1	0	1	9	0.5
Rhyolite porphyry		0	1	0	2	2	1	0	0	0	0	0	0	0	0	0	0	1	7	0.4
Argillite		0	0	0	0	0	0	0	0	0	0	0	0	0	2	0	0	0	2	0.1
Total		100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	1700	100.0

TABLE 7.—COUNTS OF PRECAMBRIAN PEBBLES FROM THE IOWAN DRIFT

Sample		12	13	76	79	95	96	97	99	Total	Average
Granite, Laurentian		27	12	21	32	17	37	30	17	193	24.1
Basic	Coarse-grained	0	0	3	3	6	0	2	5	19	2.4
	Fine-grained	8	16	23	19	6	18	5	15	110	13.8
	Fine-grained, banded	0	0	0	1	1	1	0	3	6	0.8
	Total	8	16	26	23	13	19	7	23	135	16.9
Medium basic		17	10	7	3	20	12	10	8	87	10.9
Metamorphic	Gneiss	4	8	1	1	5	3	0	1	23	2.9
	Graywacke	9	22	11	9	15	1	21	10	98	12.3
	Hornblende schist	6	1	4	11	4	7	6	10	49	6.1
	Green schist	0	1	1	0	0	0	0	1	3	0.4
	Quartzite	5	2	4	3	6	4	3	4	31	3.9
	Slate	0	0	0	0	0	3	0	3	6	0.8
	Total	24	34	21	24	30	18	30	29	210	26.2
Greenstone		12	16	15	4	12	10	20	16	105	13.1
Vein quartz		3	7	2	6	8	3	0	6	35	4.4
Rhyolite		3	5	1	2	0	0	3	0	14	1.8
Rhyolite porphyry		4	0	3	4	0	0	0	0	11	1.4
Jasper		1	0	0	2	0	1	0	0	4	0.5
Keweenawan clastics		1	0	4	0	0	0	0	1	6	0.8
Total		100	100	100	100	100	100	100	100	800	100.0

TABLE 8.—PEBBLE COUNTS FROM THE ERIE LOBE

	Sample	22	23	24	25	26	27	28	29	30	31	32	21	Total	Average
Lime-stone	Ordovician	42	38	46	46	42	51	16	17	20	29	44	39	430	35.8
	Devonian (?)	9	8	6	9	3	8	14	0	6	1	0	11	75	6.3
	Total	51	46	52	55	45	59	30	17	26	30	44	50	505	42.1
Dolomite		19	16	13	11	23	14	10	52	30	41	22	13	264	22.0
Chert		0	2	2	0	2	4	2	12	4	2	2	2	34	2.8
Siliceous coral		0	1	0	0	0	0	0	0	1	0	0	0	2	0.2
Clastic	New Albany Shale	0	0	0	0	0	0	7	0	4	0	0	0	11	0.9
	Siltstone, Devonian	11	1	13	12	8	3	16	2	12	5	2	1	86	7.2
	Sandstone, Pennsylvanian (?)	0	5	0	0	3	1	0	0	0	0	0	0	9	0.8
	Total	11	6	13	12	11	4	23	2	16	5	2	1	106	8.8
Precambrian	Acidic	4	10	3	5	9	6	7	4	2	5	3	9	67	5.6
	Basic	4	16	7	8	1	4	8	5	2	8	12	15	90	7.5
	Medium basic	4	1	3	4	5	4	7	4	7	3	6	5	53	4.4
	Metamorphic	7	2	7	5	4	4	11	3	8	6	8	3	68	5.7
	Other	0	0	0	0	0	1	2	1	4	0	1	2	11	0.9
	Total	19	29	20	22	19	19	35	17	23	22	30	34	289	24.1
Total		100	100	100	100	100	100	100	100	100	100	100	100	1200	100.0

TABLE 9.—PEBBLE COUNTS FROM THE SAGINAW LOBE

	Sample	20	33	36	37	38	42	43	44	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	Total	Average
Lime-stone	Ordovician	8	6	19	13	9	12	9	12	17	15	5	21	7	54	36	17	10	9	12	2	34	14	10	17	368	15.3
	Carboniferous	9	0	4	14	1	8	0	5	2	4	4	9	2	1	4	3	3	2	0	12	1	7	0	8	103	4.3
	Total	17	6	23	27	10	20	9	17	19	19	9	30	9	55	40	20	13	11	12	14	35	21	10	25	471	19.6
Dolo-mite	Niagara	15	8	0	0	0	0	6	3	0	0	4	4	11	0	4	7	13	5	2	2	6	13	4	6	113	4.7
	Brown, dense	20	13	24	23	9	29	22	31	34	40	7	20	9	21	9	14	10	21	33	25	12	0	5	10	441	18.5
	Total	35	21	24	23	9	29	28	34	34	40	11	24	20	21	13	21	23	26	35	27	18	13	9	16	554	23.1
Chert		1	1	7	7	2	0	1	3	0	2	5	1	3	6	5	3	5	4	0	6	1	3	5	5	76	3.2
Clastic	Sandstone, Marshall (?)	0	0	4	0	2	7	3	2	3	0	6	3	9	0	0	3	8	12	5	0	4	4	4	7	86	3.6
	New Albany Shale	0	0	0	0	0	0	0	0	0	0	0	0	0	0	1	0	0	0	0	0	0	0	0	0	1	Tr
	Siltstone, Devonian (?)	6	50	0	0	0	0	0	0	0	0	2	3	0	0	0	0	0	2	0	1	0	0	0	0	64	2.7
	Other	0	1	6	6	15	2	4	5	0	0	21	17	30	1	6	7	9	12	6	5	5	16	27	7	208	8.7
	Total	6	51	10	6	17	9	7	7	3	0	29	23	39	1	7	10	17	26	11	6	9	20	31	14	359	15.0
Precambrian	Acidic	7	5	1	5	9	6	12	6	6	11	10	6	6	2	2	8	9	0	3	9	9	12	9	7	160	6.7
	Basic	13	2	12	6	12	9	15	10	10	6	13	7	6	4	6	6	6	5	18	10	14	9	4	7	210	8.7
	Medium basic	8	3	7	6	13	7	10	3	7	9	10	4	5	0	6	13	10	9	4	6	2	3	7	9	161	6.7
	Metamorphic	12	11	15	18	24	19	17	18	18	13	12	5	11	6	17	16	13	17	16	18	11	14	21	16	358	14.9
	Other	1	0	1	2	4	1	1	2	3	0	1	0	1	5	4	3	4	2	1	4	1	5	4	1	51	2.1
	Total	41	21	36	37	62	42	55	39	44	39	46	22	29	17	35	46	42	33	42	47	37	43	45	40	940	39.1
Total		100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	2400	100.0

TABLE 10.—PEBBLE COUNTS FROM THE LAKE MICHIGAN LOBE

	Sample	1	2	3	4	5	6	7	8	9	10	14	34	35	Total	Average
Lime-stone	Ordovician	0	3	2	0	1	2	0	0	1	0	5	8	0	22	1.7
	Devonian (?)	0	0	1	0	1	0	5	1	0	0	1	0	1	10	0.8
	Total	0	3	3	0	2	2	5	1	1	0	6	8	1	32	2.5
Dolomite		18	23	41	6	75	41	33	28	8	8	18	35	7	341	26.2
Chert		4	2	2	2	3	2	1	0	1	3	3	0	2	25	1.9
Clastic	New Albany Shale	0	2	19	0	0	0	3	0	0	0	35	0	0	59	4.5
	Siltstone, Devonian	76	60	29	74	14	48	44	61	85	71	24	49	69	704	54.2
	Total	76	62	48	74	14	48	47	61	85	71	59	49	69	763	58.7
Precambrian	Acidic	0	5	1	5	2	2	2	0	0	4	5	2	5	33	2.5
	Basic	0	0	2	3	0	1	6	7	3	3	5	1	2	33	2.5
	Medium basic	2	2	0	0	4	0	1	0	0	3	2	4	5	23	1.8
	Metamorphic	0	3	0	9	0	3	5	3	1	6	1	1	8	40	3.1
	Other	0	0	3	1	0	1	0	0	1	2	1	0	1	10	0.8
	Total	2	10	6	18	6	7	14	10	5	18	14	8	21	139	10.7
Total		100	100	100	100	100	100	100	100	100	100	100	100	100	1300	100.0

TABLE 11.—PEBBLE COUNTS FROM THE GREEN BAY LOBE

	Sample	63	64	65	66	67	68	69	70	71	72	73	74	Total	Average
Lime-stone	Ordovician	5	1	5	2	0	0	0	0	0	1	0	1	15	1.3
	Other	2	0	0	0	0	0	0	0	0	0	0	0	2	0.2
	Total	7	1	5	2	0	0	0	0	0	1	0	1	17	1.4
Dolomite		76	82	72	84	93	90	88	82	87	79	77	79	989	82.4
Chert		3	2	9	1	0	0	0	0	0	4	7	10	36	3.0
Clastic	Sandstone, Cambrian (?)	1	0	0	0	0	0	1	0	0	4	2	1	9	0.8
	Siltstone, Ordovician (?)	0	0	2	1	0	0	0	0	0	0	2	0	5	0.4
	Total	1	0	2	1	0	0	1	0	0	4	4	1	14	1.2
Precambrian	Acidic	1	5	4	3	0	1	1	5	0	5	4	2	31	2.6
	Basic	2	4	1	6	3	5	4	11	8	0	1	0	45	3.8
	Medium basic	4	1	3	1	0	1	0	0	1	0	1	1	13	1.1
	Metamorphic	1	3	3	0	1	0	2	0	2	7	1	2	22	1.8
	Other	5	2	1	2	3	3	4	2	2	0	5	4	33	2.8
	Total	13	15	12	12	7	10	11	18	13	12	12	9	144	12.0
Total		100	100	100	100	100	100	100	100	100	100	100	100	1200	100.0

TABLE 12.—PEBBLE COUNTS FROM THE DES MOINES LOBE

Sample		77	78	79	80	81	82	83	84	85	86	87	88	89	90	91	92	93	Total	Average
Limestone	Mississippian	15	20	7	13	22	12	15	25	13	15	9	22	24	23	31	38	22	326	19.2
	Other	4	2	6	3	4	2	3	2	2	3	3	2	0	0	0	2	0	38	2.2
	Total	19	22	13	16	26	14	18	27	15	18	12	24	24	23	31	40	22	364	21.4
Dolomite		43	40	39	30	38	41	24	35	56	31	42	48	24	32	50	24	48	645	37.9
Chert		0	0	0	0	0	2	1	0	2	0	0	0	0	2	1	1	0	9	0.5
Clastic	Shale, Cretaceous	3	3	4	2	1	2	8	15	1	0	2	3	3	18	1	3	3	72	4.2
	Iron concretions, Cretaceous	0	1	1	1	0	0	0	0	0	0	1	1	1	0	1	0	0	7	0.4
	Carboniferous	0	1	0	0	0	1	2	0	0	1	2	0	1	3	0	3	1	15	0.9
	Total	3	5	5	3	1	3	10	15	1	1	5	4	5	21	2	6	4	94	5.5
Precambrian	Acidic	12	10	18	13	10	12	13	9	13	18	15	12	11	8	6	8	9	197	11.6
	Basic	8	8	4	18	8	12	13	8	2	17	8	3	10	4	1	7	2	133	7.8
	Medium basic	3	1	2	1	0	2	0	1	1	0	4	2	6	1	0	0	2	26	1.5
	Metamorphic	9	13	16	15	15	12	15	5	7	14	10	6	13	7	8	13	12	196	11.5
	Other	3	1	3	4	2	2	6	0	3	1	4	1	1	2	1	1	1	36	2.1
	Total	35	33	43	51	35	40	47	23	26	50	41	24	47	22	16	29	26	588	34.6
Total		100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	100	1700	100.0

TABLE 13.—PEBBLE COUNTS FROM THE IOWAN DRIFT

	Sample	12	76	79	95	96	97	99	Total	Average
Limestone	Mississippian	2	7	18	4	5	0	13	49	7.0
	Devonian; buff, porous	20	15	0	11	4	11	4	65	9.3
	Devonian; gray, platy	0	0	1	1	0	6	2	10	1.4
	Total	22	22	19	16	9	17	19	124	17.7
Dolomite		53	12	51	3	4	26	4	153	21.9
Chert		7	6	0	0	1	1	1	16	2.3
Clastic	Shale, Cretaceous	0	4	3	1	3	0	0	11	1.6
	Iron concretions, Cretaceous	0	3	0	0	0	4	1	8	1.1
	Paleozoic; sandstone and siltstone	0	3	0	1	1	1	2	8	1.1
	Total	0	10	3	2	4	5	3	27	3.9
Precambrian	Acidic	9	11	13	30	26	17	31	137	19.6
	Basic	3	10	9	5	21	4	14	66	9.4
	Medium basic	1	10	1	15	8	7	4	46	6.6
	Metamorphic	3	14	4	23	18	18	15	95	13.5
	Other	2	5	0	6	9	5	9	36	5.1
	Total	18	50	27	79	82	51	73	380	54.2
Total		100	100	100	100	100	100	100	700	100.0

TABLE 14.—SAND-GRAIN COUNTS FROM THE ERIE LOBE

	Paleozoic				Precambrian				Quartz-Feldspar			
Sample	Limestone	Dolomite	Clastic	Total*	Acidic	Basic	Metamorphic	Total	Angular	Rounded	Total	Total
22	26	14	26	66	2	1	6	9	23	2	25	100
23	37	4	31	72	0	4	0	4	24	0	24	100
24	20	9	34	64	1	6	9	16	20	0	20	100
25	18	2	48	70	0	6	1	7	23	0	23	100
26	18	5	45	69	2	1	0	3	28	0	28	100
27	29	11	34	75	1	4	2	7	17	1	18	100
28	24	7	34	65	3	5	3	11	22	2	24	100
29	31	11	35	77	1	3	8	12	11	0	11	100
30	18	20	25	63	2	3	2	7	28	2	30	100
31	23	28	16	67	0	2	0	2	29	2	31	100
32	25	19	13	58	1	3	1	5	36	1	37	100
21	36	17	3	56	3	3	3	9	31	4	35	100
Average	25	12	29	67	1	4	3	8	24	1	25	100

*includes chert

TABLE 15.—SAND-GRAIN COUNTS FROM THE SAGINAW LOBE

Sample	Paleozoic				Precambrian				Quartz-Feldspar			Total
	Limestone	Dolomite	Clastic	Total*	Acidic	Basic	Metamorphic	Total	Angular	Rounded	Total	
20	8	10	15	37	2	5	14	21	39	3	42	100
33	9	26	25	60	1	2	5	8	29	3	32	100
36	4	8	10	22	1	4	1	6	59	13	72	100
37	19	9	0	29	1	4	3	8	59	4	63	100
38	4	1	7	16	0	0	10	10	56	18	74	100
42	5	4	3	12	0	10	9	19	64	5	69	100
43	11	14	5	30	1	2	2	5	58	7	65	100
44	11	7	1	21	0	4	5	9	53	17	70	100
46	16	7	11	38	5	4	4	13	49	0	49	100
47	13	10	3	35	1	6	5	12	50	3	53	100
48	16	11	10	37	0	4	0	4	50	9	59	100
49	18	10	20	48	3	6	2	11	37	4	41	100
50	14	15	17	46	5	2	0	7	41	6	47	100
51	18	11	4	34	4	6	0	10	46	10	56	100
52	18	11	5	34	0	1	6	7	47	12	59	100
53	13	15	12	41	2	5	6	13	40	6	46	100
54	14	18	11	44	2	6	1	9	43	4	47	100
55	11	11	15	38	4	5	2	11	35	16	51	100
56	9	8	7	24	1	4	7	12	56	8	64	100
57	9	17	13	41	4	6	3	13	42	4	46	100
58	13	13	3	30	5	3	4	12	41	17	58	100
59	18	7	5	31	1	5	5	11	49	9	58	100
61	15	3	18	43	0	9	3	12	42	3	45	100
Average	12	11	10	34	2	5	4	11	47	8	55	100

*includes chert

TABLE 16.—SAND-GRAIN COUNTS FROM THE LAKE MICHIGAN AND GREEN BAY LOBES

Sample	Paleozoic				Precambrian				Quartz-Feldspar			Total
	Limestone	Dolomite	Clastic	Total*	Acidic	Basic	Metamorphic	Total	Angular	Rounded	Total	
Lake Michigan Lobe												
1	7	26	60	93	1	0	1	2	4	1	5	100
2	6	21	62	89	0	1	2	3	4	4	8	100
3	0	32	51	83	0	3	3	6	10	1	11	100
4	6	17	65	88	0	1	4	5	5	2	7	100
5	4	63	26	93	0	1	2	3	3	1	4	100
6	6	45	28	79	1	0	2	3	13	5	18	100
7	25	23	41	89	0	0	1	1	5	5	10	100
8	21	26	43	90	0	1	1	2	3	5	8	100
9	14	30	42	86	3	1	3	7	5	2	7	100
10	4	31	49	84	0	1	0	1	10	5	15	100
34	8	19	39	66	0	1	10	11	15	8	23	100
35	5	19	48	72	0	3	4	7	13	8	21	100
Average	9	29	46	85	Tr	1	3	4	8	3	11	100
Green Bay Lobe												
63	1	31	2	37	1	6	18	25	22	16	38	100
64	0	16	2	23	3	0	3	6	29	42	71	100
65	1	17	5	24	6	2	3	11	34	31	65	100
66	3	33	16	54	0	1	2	3	28	15	43	100
67	1	60	4	68	0	5	2	7	19	6	25	100
68	0	81	0	81	2	2	7	11	7	1	8	100
69	0	55	12	67	1	2	14	17	14	2	16	100
70	2	30	11	44	3	2	12	17	22	17	39	100
71	0	27	0	28	2	2	10	14	19	39	58	100
72	0	32	1	33	5	0	8	13	27	27	54	100
73	2	11	1	20	2	1	4	7	61	12	73	100
74	0	16	2	19	1	24	13	38	36	7	43	100
Average	1	34	5	42	2	4	8	14	26	18	44	100

*includes chert

TABLE 17.—SAND-GRAIN COUNTS FROM THE DES MOINES LOBE AND IOWAN DRIFT

Des Moines Lobe												
	Paleozoic and Mesozoic				Precambrian				Quartz-Feldspar			
Sample	Limestone	Dolomite	Clastic	Total*	Acidic	Basic	Metamorphic	Total	Angular	Rounded	Total	Total
77	5	3	10	20	1	10	2	13	59	8	67	100
78	10	3	11	25	5	2	2	9	57	9	66	100
79	3	9	15	29	3	3	3	9	48	14	62	100
80	7	5	11	24	8	0	1	9	61	6	67	100
81	10	7	12	30	3	2	2	7	57	6	63	100
82	7	5	18	33	4	1	0	5	55	7	62	100
83	8	4	22	34	1	2	2	5	54	7	61	100
84	12	6	34	53	2	3	0	5	37	5	42	100
85	8	11	15	34	4	4	3	11	51	4	55	100
86	11	12	6	30	3	2	1	6	59	5	64	100
87	8	11	17	37	1	3	2	6	50	7	57	100
88	3	12	20	35	2	2	1	5	50	10	60	100
89	1	8	33	43	1	1	1	3	49	5	54	100
90	8	10	25	43	2	2	2	6	45	6	51	100
91	7	9	42	58	2	2	0	4	33	5	38	100
92	5	6	44	57	4	3	0	7	30	6	36	100
93	9	10	13	33	4	4	2	10	46	11	57	100
Average	7	8	20	36	3	3	1	7	50	7	57	100
					Iowan Drift							
12	4	14	0	21	0	1	1	2	47	30	77	100
76	6	4	3	13	1	1	3	5	73	9	82	100
94	2	0	3	5	0	0	2	2	70	23	93	100
95	15	0	6	22	0	2	4	6	69	3	72	100
96	2	1	3	7	0	4	0	4	78	11	89	100
97	5	6	5	16	0	2	2	4	71	9	80	100
99	1	1	1	3	1	4	1	6	70	21	91	100
Average	5	4	3	12	Tr	2	2	4	69	15	84	100

*includes chert

TABLE 18.—PEBBLE COUNTS FROM KANSAN TILL

Sample		K1	K2	K3	K4	K5	K6	Total	Average
Limestone	Mississippian	10	8	2	19	14	8	61	10.1
	Devonian (?)	11	1	1	4	2	7	26	4.3
	Total	21	9	3	23	16	15	87	14.5
Dolomite		13	50	23	10	8	8	112	18.7
Chert		6	14	20	6	20	0	66	11.0
Clastics		1	4	17	0	7	33	62	10.3
Iron concretions		0	0	1	1	0	0	2	0.3
Precambrian	Acidic	15	9	19	25	16	19	103	17.2
	Basic	16	4	1	5	12	11	49	8.2
	Medium basic	5	2	1	4	3	0	15	2.5
	Metamorphic	21	5	11	20	16	13	86	14.3
	Other	2	3	4	6	2	1	18	3.0
	Total	59	23	36	60	49	44	271	45.1
Total		100	100	100	100	100	100	600	100.0

TABLE 19.—COUNTS OF PRECAMBRIAN PEBBLES FROM KANSAN TILL

Sample		K1	K2	K3	K4	K5	K6	Total	Average
Granite, Laurentian		12	25	13	22	18	24	114	19.0
Basic	Coarse-grained	8	4	8	4	5	4	33	5.5
	Fine-grained	12	8	11	14	20	28	93	15.5
	Fine-grained, banded	0	0	2	0	1	0	3	0.5
	Total	20	12	21	18	26	32	129	21.5
Medium basic		14	13	5	10	6	4	52	8.7
Metamorphic	Gneiss	0	5	0	5	1	1	12	2.0
	Graywacke	12	9	29	22	21	12	105	17.5
	Hornblende schist	11	11	4	5	6	9	46	7.7
	Green schist	0	5	1	2	0	1	9	1.5
	Quartzite	7	0	5	8	10	3	33	5.5
	Slate	1	0	0	1	1	0	3	0.5
	Total	31	30	39	43	39	26	208	34.7
Greenstone		11	12	9	1	4	4	41	6.8
Vein quartz		4	5	8	5	5	9	36	6.0
Rhyolite		2	2	4	0	1	0	9	1.5
Rhyolite porphyry		2	0	0	1	1	1	5	0.8
Keweenawan clastic		1	0	0	0	0	0	1	0.2
Jasper		3	1	1	0	0	0	5	0.8
Total		100	100	100	100	100	100	600	100.0

TABLE 20.—DESCRIPTION OF SAMPLE LOCATIONS

Sample number	Location	Lobe	Moraine
1	Railroad cut 1 mile southwest of Matteson, Ill., Rt. 54; NE ¼, NW ¼, Sec. 35, T. 35 N., R. 13 E.	Lake Michigan	Tinley
2	Railroad cut 2.5 miles southwest of Matteson, Ill., Rt. 54, NE ¼, SW ¼, Sec. 3, T. 34 N., R. 13 E.	Lake Michigan	Zurich
3	Road cut 2 miles east of Monee, Ill., NE corner Sec. 29, T. 34 N., R. 13 E.	Lake Michigan	Monee
4	Railroad cut 2.5 miles southwest of Monee Ill., Rt. 54, SW ¼, Sec. 32, T. 34 N., R. 13 E.	Lake Michigan	West Chicago
5	Road cut 1 mile east of Whitaker, Ill., SW ¼, Sec. 14, T. 32 N., R. 13 E.	Lake Michigan	Rockdale
6	Road cut 4 miles south of Eagle Lake, Ill., NE ¼, Sec. 28, T. 33 N., R. 14 E.	Lake Michigan	Rockdale (?), Monee (?)
7	Road cut 2 miles east of Steger, Ill., SW ¼. SW ¼, Sec. 28, T. 35 N., R. 14 E.	Lake Michigan	Tinley
8	Road cut 9 miles south of Steger, Ill., center of west side Sec. 21, T. 34 N., R. 14 E.	Lake Michigan	Zurich
9	Road cut 1 mile west of Beecher, Ill., SE ¼, Sec. 18, T. 33 N., R. 14 E.	Lake Michigan	Monee
10	Road cut 1 mile east of Sollitt, Ill., NE ¼, Sec. 3, T. 32 N., R. 14 E.	Lake Michigan	West Chicago
11	Road cut 3.5 miles southeast of Grant Park, Ill., NE ¼, Sec. 35, T. 32 N., R. 14 E.	Lake Michigan	Manhattan
12	Road cut 3.7 miles south of DeWitt, Ia., SW ¼, Sec. 31, T. 81 N., R. 4 E.	Iowan drift	(Outwash sand)
13	Road cut 8.5 miles south of Vinton, Ia., SW ¼, T. 84 N., R. 10 W.	Iowan drift	—
14	Road cut 5 miles north of Valparaiso, Ind., NW ¼, SE ¼, NW ¼, Sec. 35, T. 36 N., R. 6 W.	Lake Michigan	Valparaiso
15	Excavation, N. Riverside, Ill., SW ¼, Sec. 24, T. 39 N., R. 12 E.	Lake Michigan	Tinley (ground moraine)
16	Road cut 6 miles north of Valparaiso, Ind., SE ¼, Sec. 24, T. 36 N., R. 6 E.	Lake Michigan	Valparaiso
17	Excavation, Walkerton, Ind., NW ¼, Sec. 23, T. 35 N., R. 1 W.	Saginaw	Maxinkuckee (outwash)
18	Road cut 1 mile west of Lapaz, Ind., NE ¼, Sec. 7, T. 35 N., R. 2 E.	Saginaw	Maxinkuckee
19	Road cut at Syracuse, Ind., center of east side of Sec. 7, T. 34 N., R. 7 E.	Saginaw	New Paris (sand)
20	Road cut 1.8 miles north of North Webster, Ind., SW ¼, Sec. 35, T. 34 N., R. 7 E.	Saginaw	New Paris
21	Road cut 1 mile north of Pierceton, Ind., SW ¼, Sec. 22, T. 32 N., R. 7 E.	Erie	Interlobate
22	Road cut 4 miles northwest of Columbia City, Ind., NW ¼, Sec. 6, T. 31 N., R. 9 E.	Erie	Mississinnawana
23	Road cut 4 miles southeast of Columbia City, Ind., SW ¼, Sec. 17, T. 31 N., R. 10 E.	Erie	Salamonie
24	Road cut 5 miles northwest of Fort Wayne, Ind., NW ¼, Sec. 19, T. 31 N., R. 12 E.	Erie	Wabash
25	Road cut 8 miles southeast of Fort Wayne, Ind., NW ¼, T. 29 N., R. 13 E.	Erie	Fort Wayne
26	Road cut 3 miles north of Decatur, Ind., SW ¼, T. 28 N., R. 14 E.	Erie	Forth Wayne
27	Road cut 11 miles west of Decatur, Ind., SE ¼, Sec. 33, T. 28 N., R. 12 E.	Erie	Wabash (ground moraine)

TABLE 20.—*Continued*

Sample number	Location	Lobe	Moraine
28	Road cut 2 miles east of Uniondale, Ind., NW ¼, Sec. 5, T. 27 N., R. 12 E.	Erie	Wabash
29	Road cut 5 miles southeast of Huntington, Ind., SE ¼, T. 28 N., R. 9 E.	Erie	Salamonie
30	Road cut 3.5 miles northwest of Wabash, Ind., SW ¼, Sec. 32, T. 28 N., R. 7 E.	Erie	Mississinnawana
31	Road cut 1 mile north of Mexico, Ind., SW ¼, T. 28 N., R. 4 E.	Erie	Interlobate
32	Road cut 6 miles southeast of Rochester, Ind., NE ¼, T. 29 N., R. 3 E.	Erie	Interlobate
33	Road cut 2 miles east of Donaldson, Ind., NE ¼, Sec. 34, T. 34 N., R. 1 E.	Saginaw	Maxinkuckee
34	Road cut at Valparaiso, Ind., NW ¼, Sec. 25, T. 35 N., R. 6 W.	Lake Michigan	Valparaiso
35	Road cut 7 miles east of Dyer, Ind., SE ¼, T. 35 N., R. 9 W.	Lake Michigan	Valparaiso
36	Road cut 7.5 miles east of Elkhart, Ind., NE ¼, T. 37 N., R. 6 E.	Saginaw	Middlebury
37	Road cut 10 miles east of Bristol, Ind., SW ¼, T. 38 N., R. 8 E.	Saginaw	La Grange (gravel)
38	Road cut 1 mile north of Coldwater, Mich., NE ¼, T. 6 S., R. 7 W.	Saginaw	Sturgis (gravel)
39	Road cut 1.5 miles northeast of Tekonsha, Mich., SE ¼, T. 4 S., R. 6 W.	Saginaw	Tekonsha (gravel)
40	Excavation 1 mile south of Albion, Mich., NW ¼, T. 3 S., R. 4 W.	Saginaw	Kalamazoo (gravel)
41	Excavation 1 mile south of Springport, Mich., SW ¼, T. 1 S., R. 3 W.	Saginaw	Kalmaazoo (gravel)
42	Road cut 4.5 miles south of Eaton Rapids, Mich., SE ¼, T. 1 N., R. 3 W.	Saginaw	Kalamazoo
43	Road cut 1 mile northwest of Eaton Rapids, Mich., SW ¼, T. 2 N., R. 3 W.	Saginaw	Charlotte
44	Road cut 1 mile northeast of Charlotte, Mich., NW ¼, T. 2 N., R. 4 W.	Saginaw	Charlotte
45	Road cut at Millett, Mich., SE ¼, T. 4 N., R. 3 W.	Saginaw	Lansing
46	Excavation 2 miles northeast of Lansing, Mich., NE ¼, T. 4 N., R. 2 W.	Saginaw	Grand Ledge
47	Road cut 2 miles southwest of Schaftsburg, Mich., SW ¼, T. 5 N., R 1 E.	Saginaw	Ionia
48	Road cut 1 mile northeast of Bancroft, Mich., SE ¼, T. 6 N., R. 3 E.	Saginaw	Portland
49	Road cut 4 miles northwest of Durand, Mich., NW ¼, T. 7 N., R. 5 E.	Saginaw	Flint
50	Road cut at Lennon, Mich., SW ¼, T. 7 N., R. 5 E.	Saginaw	Owosso
51	Excavation at Saginaw, Mich., SW ¼, T. 12 N., R. 5 E.	Saginaw	Port Huron
52	Excavation at Saginaw, Mich, SW ¼, T. 12 N., R 4 E.	Saginaw	Port Huron
53	Road cut at Shepardsville, Mich., NW ¼, T. 7 N., R. 1 W.	Saginaw	St. Johns
54	Road cut 4 miles south of St. Johns, Mich., NE ¼, T. 6 N., R. 2 W.	Saginaw	Lyons
55	Road cut 8 miles north of Lansing, Mich., NE ¼, T. 5 N., R. 2 W.	Saginaw	Portland

TABLE 20.—*Continued*

Sample number	Location	Lobe	Moraine
56	Road cut 4 miles north of Lansing, Mich., SE ¼, T. 5 N., R. 2 W.	Saginaw	Ionia
57	Road cut 1 mile southeast of Lansing, Mich.. SW ¼, T. 4 N., R. 2 W.	Saginaw	Lansing
58	Road cut 8 miles north of Marshall, Mich., SE ¼, T. 1 S., R. 6 W.	Saginaw	Kalamazoo
59	Road cut 6 miles northeast of Celon, Mich., SE ¼, T. 5 S., R. 8 W.	Saginaw	Sturgis (gravel)
60	Road cut 1 mile northeast of Celon, Mich., NW ¼, T. 6 S., R. 9 W.	Saginaw	Sturgis (gravel)
61	Road cut 2 miles east of White Pigeon, Mich., NW ¼, T. 8 S., R. 11 W.	Saginaw	Sturgis (gravel)
63	Road cut 5 miles southeast of Whitewater, Wis., NW ¼, Sec. 29, T. 4 N., R. 15 E.	Green Bay	Interlobate
64	Pit 1 mile south of Jefferson, Wis., NE ¼, T. 6 N., R. 14 E.	Green Bay	Lake Mills
65	Excavation 2 miles north of Johnson Creek, Wis., NW ¼, T. 7 N., R. 15 E.	Green Bay	Lake Mills
66	Road cut 1 mile south of Lebanon, Wis., NW ¼, T. 9 N., R. 16 E.	Green Bay	Lake Mills
67	Road cut 3 miles west of Fond du Lac, Wis., NW ¼, T. 15 N., R. 17 E.	Green Bay	Valders
68	Pit 6 miles southeast of Ripon, Wis., T. 16 N., R. 15 E.	Green Bay	Rush Lake
69	Road cut 1 mile south of Brandon, Wis., SE ¼, T. 15 N., R. 15 E.	Green Bay	Green lake
70	Road cut 2 miles south of Randolph, Wis., NW ¼, T. 12 N., R. 13 E.	Green Bay	Lake Mills
71	Road cut 5 miles north of Waterloo, Wis., SW ¼, T. 9 N., R 12 E.	Green Bay	Lake Mills
72	Pit 2 miles northwest of Lake Mills, Wis., NE ¼, T. 7 N., R 13 E.	Green Bay	Lake Mills
73	Road cut 2 miles northwest of Whitewater, Wis., SW ¼, T. 5 N., R. 15 E.	Green Bay	Milton
74	Road cut 2 miles north of Richmond, Wis., NW ¼, T. 3 N., R. 15 E.	Green Bay	Johnstown
76	Road cut 1 mile west of Tama, Ia., SW ¼, T. 83 N., R. 15 W.	Iowan	—
77	Stream cut 6 miles southwest of Marshalltown, Ia., NE ¼, T. 83 N., R 19 W.	Des Moines	—
78	Road cut 6 miles west of Ames, Ia., NE ¼, T. 83 N., R. 25 W.	Des Moines	—
79	Pit at Des Moines River and U.S. 30, 3 miles southwest of Boone, Ia., SW ¼ T. 84 N., R. 26 W.	Iowan and Des Moines	—
80	Road cut 4 miles west of Ogden, Ia., SW ¼, T. 84 N., R. 28 W.	Des Moines	—
81	Road cut 6 miles north of Jefferson, Ia., NW ¼, T. 84 N., R. 30 W.	Des Moines	—
82	Stream cut 1.5 miles east of Churdan, Ia., SE ¼, T. 85 N., R. 31 W.	Des Moines	—
83	Road cut 1 mile east of Churdan, Ia., SW ¼, T. 85 N., R. 31 W.	Des Moines	—
84	Large cut at Dayton, Ia., NW ¼, T. 86 N., R. 28 W.	Des Moines	—
85	Road cut 1.6 miles north of Stanhope, Ia., SW ¼, T. 87 N., R. 25 W.	Des Moines	—

TABLE 20.—*Continued*

Sample number	Location	Lobe	Moraine
86	Stream cut 2 miles south of Webster City, Ia., NW ¼, T. 88 N., R. 25 W.	Des Moines	—
87	Road cut ½ mile west of Des Moines River at Fort Dodge, Ia., T. 89 N., R. 29 W.	Des Moines	—
88	Stream cut 4 miles north of Fort Dodge, Ia., SW ¼, T. 90 N., R. 28 W.	Des Moines	—
89	Road cut 1 mile south of Rutland, Ia., SW ¼, T. 92 N., R. 29 W.	Des Moines	—
90	Road cut at Des Moines River at Livermore, Ia., NW ¼, T. 93 N., R. 28 W.	Des Moines	—
91	Road cut 1 mile south of Algona, Ia., NE ¼, T. 95 N., R. 29 W.	Des Moines	—
92	Road cut 10 miles west of Algona, Ia., SE ¼, T. 96 N., R. 27 W.	Des Moines	—
93	Stream cut 3 miles east of Britt, Ia., SW ¼, T. 96 N., R. 25 W.	Des Moines	—
94	Road cut 1 mile south of Floyd, Ia., SW ¼, T. 96 N., R. 16 W.	Iowan drift	(gravel)
95	Road cut 2 mile east of Plainfield, Ia., SW ¼, T. 93 N., R. 14 W.	Iowan	—
96	Road cut 2 miles northwest of Washburn, Ia., NW ¼, T. 88 N., R. 12 W.	Iowan	—
97	Road cut 5 miles west of Vinton, Ia., SE ¼, T. 85 N., R. 11 W.	Iowan	—
99	Road cut 2 miles east of Mount Vernon, Ia., NW ¼, T. 82 N., R. 5 W.	Iowan	—
K1	Road cut, east bluff of Mississippi valley, NW ¼, NW ¼, sec. 30, T. 16 N., R. 5 W.	Kansan	
K2	Road cut 7 miles west of Muscatine, Ia., NW ¼, SE ¼, sec. 6, T. 76 N., R. 2 W.	Kansan	
K3	Road cut in Loud Thunder State Park, SW ¼, SE ¼, sec. 27, T. 17 N., R. 4 W.	Kansan	
K4	Road cut west of Mill Creek Valley, SW ¼, NW ¼, sec. 7, T. 16 N., R. 1 W.	Kansan	
K5	Stream cut north side of Keating Creek, SE ¼, SW ¼, sec. 8, T. 15 N., R. 5 W.	Kansan	
K6	Stream cut 2 miles southwest of Spring Valley, Ill., NW ¼, NW ¼, sec. 28, T. 33 N., R. 1 W.	Kansan	

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1451–1514, 6 FIGS., 1 PL. NOVEMBER 1957

CORRELATION OF THE TRIASSIC FORMATIONS OF NORTH AMERICA EXCLUSIVE OF CANADA

BY THE TRIASSIC SUBCOMMITTEE, JOHN B. REESIDE, JR. (*Chairman*), P. L. APPLIN, E. H. COLBERT, J. T. GREGORY, H. D. HADLEY, BERNHARD KUMMEL, P. J. LEWIS, J. D. LOVE, MANUEL MALDONADO-KOERDELL, E. D. MCKEE, D. B. MCLAUGHLIN, S. W. MULLER, J. A. REINEMUND, JOHN RODGERS, JOHN SANDERS, N. J. SILBERLING, AND KARL WAAGÉ

CONTENTS

TEXT

	Page
Introduction	1451
Classification of the Triassic System	1455
Correlation of continental Triassic sediments by vertebrate fossils. By E. H. Colbert and J. T. Gregory	1456
Areas of exposure	1456
Lower Triassic	1457
Upper Triassic	1457
General Discussion	1457
Eastern North America	1459
Wind River Mountains, Wyoming	1460
Northern Arizona	1462
Northern New Mexico	1464
Eastern New Mexico	1464
Texas	1465
Rhaetic deposits	1466
Annotations	1467
Index of stratigraphic names by columns	1502
References cited	1503

ILLUSTRATIONS

Figure	Page
1. Index map of Central America and Mexico showing areas represented by columns of Chart 8a numbered 1 to 9	1452
2. Index map of the United States showing areas represented by columns of Chart 8a numbered 10 to 94	1453
3. Index map of Pennsylvania and adjacent States showing in greater detail than in Figure 2 areas represented by columns of Chart 8a numbered 79 to 92	1454
4. Index map of Alaska showing areas represented by columns of Chart 8a numbered 95 to 109	1454
5. Index map of Greenland showing areas represented by columns of Chart 8a numbered 110 and 111	1455
6. Map of Florida and parts of adjacent States showing location of wells in which Triassic (?) rocks have been found and which are listed in Table 4	1487

Plate	Facing page
1. Correlation of Triassic formations of North America exclusive of Canada (Chart 8a of series)	1451, 1514

TABLES

Table	Page
1. Vertebrates of the Newark Group	1461
2. Vertebrates of the Upper Triassic of Wyoming, Arizona, and northern New Mexico	1463
3. Vertebrates of the Upper Triassic of east-central New Mexico, northeastern New Mexico, Colorado-Oklahoma, and Texas	1465
4. Triassic (?) rock in wells in Alabama, Georgia, and Florida	1488

INTRODUCTION

This is Number 8a of a series of correlation charts of North American sedimentary formations prepared by the Committee on Stratigraphy of the National Research Council (Dunbar, 1942). It presents a correlation of the Triassic formations of North America exclusive of Canada. Chart Number 8b, for the Triassic formations of Canada, was published some years ago (McLearn, 1953).

General papers on the Triassic stratigraphy of North America are few. Except for the brief treatments in most textbooks of historical geology, the discussion by Willis (1912, p. 500–508) is the only one. Russell in 1892 summarized what was known of the Newark Group of eastern North America, and, as noted above, McLearn brought together in 1953 the data on the Triassic of Canada. Smith (1904; 1914; 1927; 1932) dealt with the stratigraphy incidental to his studies of the marine Triassic faunas. Most other contributions concern only restricted areas.

The Triassic rocks of the westernmost part of North America are chiefly marine deposits associated with much volcanic material. These pass eastward into nonmarine red beds in the

Plateau and Rocky Mountain regions of the United States. The central part of the continent has scant or no Triassic deposits, but in the easternmost part mostly reddish nonmarine

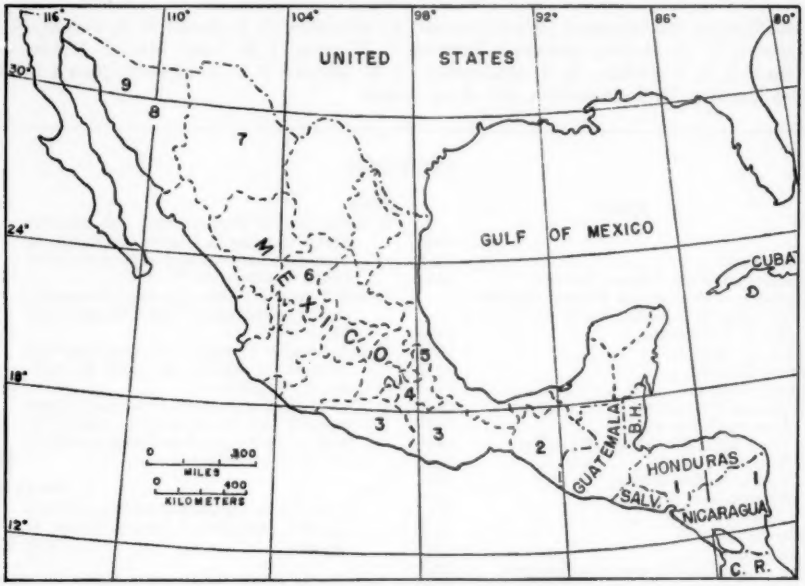

FIGURE 1.—INDEX MAP OF CENTRAL AMERICA AND MEXICO SHOWING AREAS REPRESENTED BY COLUMNS OF CHART 8A NUMBERED 1 TO 9

deposits associated locally with dark volcanic rocks are interpreted to be of Late Triassic age and extend from North Carolina to New England. Rocks found in wells in the southeastern United States are construed as probably an extension of these Late Triassic deposits.

The marine deposits, because of the cosmopolitan character of the marine faunas, lend themselves readily to correlation within the continent and with deposits on other continents. The nonmarine deposits, on the other hand, offer many more difficulties and require much more expression of personal opinion and interpretation than do the marine deposits. The nonmarine deposits are necessarily provincial and the assignment of local deposits to a standard scale of stages entails consideration of physical characteristics of the sediments and the biological content as well. In general, the nonmarine fauna and the flora are not abundantly distributed through the sequence of rocks, and opinions of well-informed persons may differ as to the meaning of the available fossils in terms of a general scale.

The chart, Plate 1, contains numbered columns representing the Triassic sequence in 111 areas, arranged by States or other large political divisions. The general order of arrangement is intended to be from south to north and from west to east, although departures from this order will be found. Each column includes the stratigraphic units recognized in the area represented and bears the compiler's name. Some 150 stratigraphic names, of various ranks, are noted in the chart, and an index of them referred to column numbers is given below. (*See* Index by columns of stratigraphic names.) Some of the names used in earlier publications have been displaced at various times and for various reasons and are not considered here. Most of those used up to 1936 may be found in Wilmarth's *Lexicon of geologic names* (1938).

The index maps show by column number the general location of each area represented by a column in Central America and Mexico (Fig. 1), the United States (Fig. 2), in more detail in Pennsylvania and adjacent States (Fig. 3), Alaska (Fig. 4), and Greenland (Fig. 5).

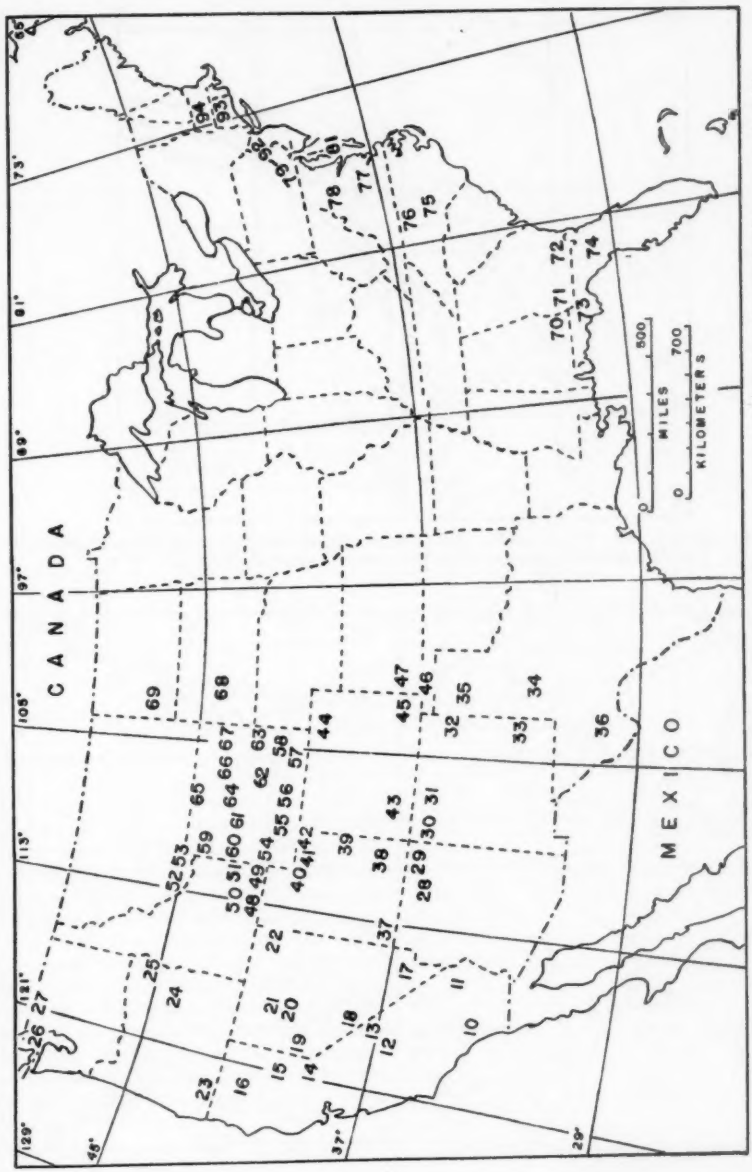

FIGURE 2.—INDEX MAP OF THE UNITED STATES SHOWING AREAS REPRESENTED BY COLUMNS OF CHART 8A NUMBERED 10 TO 94

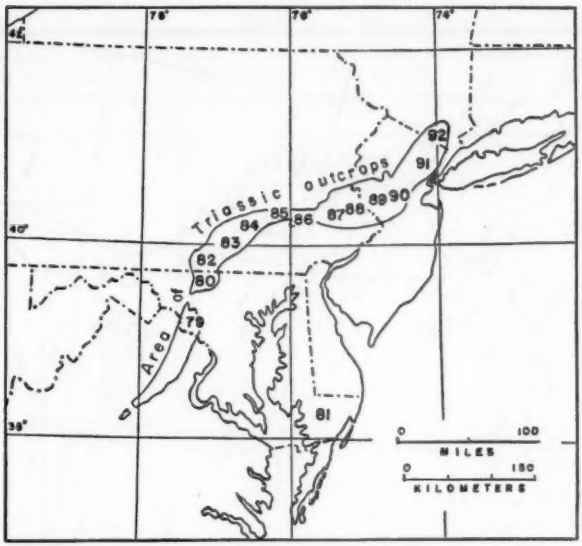

FIGURE 3.—INDEX MAP OF PENNSYLVANIA AND ADJACENT STATES SHOWING IN GREATER DETAIL THAN IN FIGURE 2 AREAS REPRESENTED BY COLUMNS OF CHART 8A NUMBERED 79 TO 92

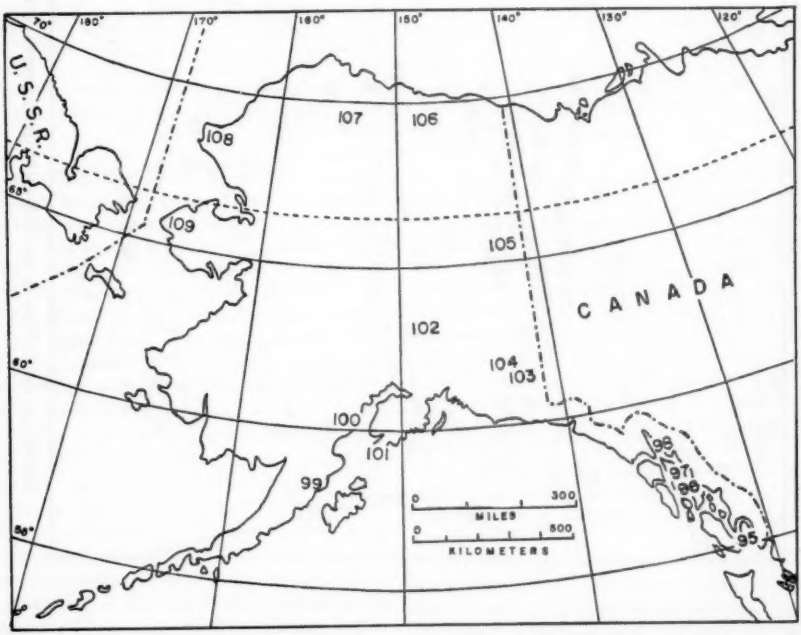

FIGURE 4.—INDEX MAP OF ALASKA SHOWING AREAS REPRESENTED BY COLUMNS OF CHART 8A NUMBERED 95 TO 109

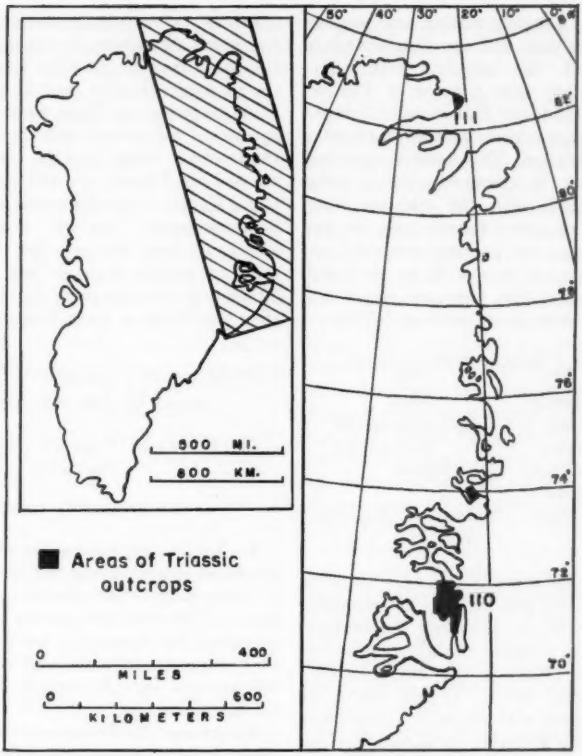

FIGURE 5.—INDEX MAP OF GREENLAND SHOWING AREAS REPRESENTED BY COLUMNS OF CHART 8A NUMBERED 110 AND 111

Chronologic relations are indicated in the chart by horizontal position—*i. e.*, the same level indicates approximately the same age. Hiatuses are indicated by vertical lining; lack of knowledge is indicated by diagonal lining. Numbers within the columns of the chart refer to annotations given below (*see* Annotations), and each annotation is signed by the contributor. The chief sources of data are shown by numbers at the bottom of each column referred to the list of publications given below. (*See* References cited.) Abbreviations used in the chart are as follows:

blk.	black
calc.	calcareous
can., c.	canyon
conglom., cgl.	conglomerate
fangl., fgl.	fanglomerate
fl.	flow
form., f.	formation
low.	lower
ls.	limestone
memb., m.	member
mid.	middle
pt.	part
sandst., sand., sst., ss.	sandstone
sh.	shale
siltst.	siltstone
sp., spp.	species, singular and plural
tong., t.	tongue
up.	upper

CLASSIFICATION OF THE TRIASSIC SYSTEM

The Triassic System, or Trias, derives its name from a sequence of rocks in western Europe in which three major divisions are recognized. This sequence contains a lower

nonmarine unit, a middle marine unit, and an upper nonmarine unit, and constitutes what is commonly called "the Germanic type." For years this was the only concept of Triassic rocks. Then it was found that rocks of Triassic age in the Alps are entirely marine and need a different classification. This marine sequence constitutes what is commonly called "the Alpine type" and, with the addition from southern Asia of some minor units in the lower beds, forms the primary standard sequence of stages. A concensus as to broad correlation between the Germanic type and the Alpine type may be set down as follows:

Series	Stages of Alpine type		Stages of Germanic type
Upper	Rhaetian, Norian, Carnian	Keuper	Rhät, Gypskeuper, Lettenkohle
Middle	Ladinian, Anisian	Muschelkalk	
Lower	Scythian	Bunter or Buntsandstein	

Departures from this concensus are found in the literature. Some geologists, as, for example, Arkell (1933, p. 97) in discussing the Jurassic of Great Britain, consider the Rhaetian to be Jurassic. Spath (1934, p. 20, 39), however, considered certain British Liassic strata to have been confused with strata of Triassic age and accepted the Rhaetian as distinct and as Triassic. In a recent publication Arkell (1956, p. 172) seems to accept the Rhaetian as Triassic. French geologists (*e. g.*, Gignoux, 1955, p. 265, 270, 314) define the term Keuper in more restricted fashion than do the German geologists, apply Keuper to the middle part of the German sequence (Gypskeuper), and refer the Rhaetian (Rhät) to the Jurassic and the Lettenkohle to the Middle Triassic (Muschelkalk). In the French literature the German terms have been translated, in ascending order, as *Gres bigarré* (Bunter), *Calcaire conchylien* (Muschelkalk), and *Marnes irisées* (Keuper, restricted). An additional but minor variation is the use of Virglorian for Anisian and Werfenian for Scythian.

In the chart (Pl. 1) the marine stages and a set of faunal zones derived from the Alpine type of Triassic are given as the standard divisions, and an effort is made to correlate the American units with this standard. Necessarily much guessing of details is required, though some faunal zones are widespread. For the nonmarine parts of the section there is considerable agreement on internal correlation but much divergence on correlation with the Germanic type of Triassic. Even here a concensus seems to exist, and an effort has been made to follow it. There seems to be general agreement, for example that the Newark Group, the Dockum Group, and the Chinle Formation are of Late Triassic age and about equivalent to the Keuper of the Germanic type of Triassic. Some geologists, however, would extend the length of time covered by these American units to include more or less of the Middle Triassic of the Germanic type. In the chart they are shown as Late Triassic.

Correlation of Continental Triassic Sediments by Vertebrate Fossils

By Edwin H. Colbert and Joseph T. Gregory

Areas of Exposure

In North America, continental Triassic deposits are exposed along the Atlantic Coast and in the Rocky Mountain-Colorado Plateau region. The western exposures are only slightly obscured by vegetation or other cover, in contrast to those in the east which are generally covered by soil, vegetation, or the works of man.

In the east the Triassic continental deposits, which occur in a series of discontinuous basins from New Brunswick on the north to North Carolina and South Carolina at the south, are called the Newark Group, except in New Brunswick where they are termed the Quaco Conglomerate. In the Colorado Plateau area Lower Triassic deposits are termed the Moenkopi Formation; to the west and north these grade into Thaynes Formation and associated marine deposits. Unconformably overlying the Moenkopi and overlapping it extensively to the east is the Chinle Formation, with a basal sandstone or conglomerate often separately recognized as the Shinarump. In eastern New Mexico and western Texas, equivalents of the Chinle are termed the Dockum Group. An outlier along the northwest flank of the Glass Mountains is called the Bissett Conglomerate.

From the Wind River Mountains of Wyoming east into South Dakota extensive Triassic red beds are designated as the Chugwater and Spearfish Formations. These extend downward to include Permian strata in the east, but to the west the marine Dinwoody Formation of

Early Triassic age separates them from the underlying Permian beds. The upper member of the Chugwater, the Popo Agie, is typically continental and contains a Late Triassic fauna similar to that of the Dockum and Chinle.

Lower Triassic

The Moenkopi Formation of Arizona and Utah consists of dark-red sandstones and siltstones with interbedded gypsum in the upper part and marine limestones in the lower portion of its more western exposures. It thins and becomes entirely continental to the east; to the west it passes into the marine Triassic deposits of the Cordilleran geosyncline. An Early Triassic age is indicated by marine invertebrates in the limestone members, but, as noted below, the vertebrate evidence indicates a possible extension of the Moenkopi into Middle Triassic time.

In eastern Arizona the continental facies is divided into three members, named from below the Wupatki, Moqui, and Holbrook.

Fossil vertebrates occur in both the lower (Wupatki) and upper (Holbrook) members of the Moenkopi (Welles, 1947; Peabody, 1948). The Moqui Member is barren and apparently was unfavorable for terrestrial organisms.

Localities at Meteor Crater and Moqui Wash, west of Winslow, have produced the following fauna from the lower part of the Moenkopi: Capitosaurid tracks, *Chirotherium*, *Rotodactylus*, *?Aphaneramma*, Capitosaurids (cf. *Parotosaurus*), and *Stanocephalosaurus*.

Localities in the upper part of the Moenkopi at Holbrook (Penzance) and Cameron, Arizona, contain: *Anisodontosaurus*, *Arizonasaurus*, *Chirotherium barthi*, *Taphrognathus*, *Cyclotosaurus*, *Rhadalognathus*, [Brachyopid, near Cameron, not published], *Leiacanthus*, *Boreosomus*, dipnoans, coelacanths.

The lower fauna resembles that of the Early Triassic of Spitzbergen and the Bunter of Germany. The fauna of the upper part of the Moenkopi Formation is generally of Lower Triassic aspect. But the advanced amphibian genus *Cyclotosaurus* suggests possible Middle Triassic age. *Chirotherium* tracks have been found at other Moenkopi localities as far north as Vernal, Utah.

Footprints (*Collettosaurus*, *Eurichnus*) have been found in the Red Peak Member in the lower part of the Chugwater of Wyoming (Lull, 1942). These differ from those of the Moenkopi Formation, but their stratigraphic position below the Alcova Limestone indicates Early Triassic age.

Upper Triassic

General discussion—A comparison of faunal lists for the Newark Group, Chinle Formation, Dockum Group, and Popo Agie Member of the Chugwater Formation might easily lead an uncritical student to think that great differences distinguish the several faunas known from these beds. Much of the confusion about these faunas is due to the manner in which they have been studied and described over the years. The first fossil vertebrates from continental Triassic sediments in North America were described from the Newark Group. The remains were for the most part rare and fragmentary fossils, and the knowledge to be gained from them was, and frequently still is, far from satisfactory. During the past half century several students have given their attention to fossils from the western Triassic beds. Yet these authorities for the most part have worked at the problem piecemeal. One man would devote his attention to fossils from western Texas, another to those from Arizona, a third to those from Wyoming. Consequently there were not so many broad comparisons or syntheses as might have been desirable. And since good specimens are not easy to acquire, even in the comparatively productive beds of the southwest, there was an understandable tendency to consider many a skull or skeleton as representative of a new genus or, at the least, of a new species. Other factors complicated the problem. Some of the types, because they were the first specimens discovered, were based upon fragmentary remains. This made it difficult to compare subsequent specimens with the types, so that fossils collected at later dates were often given new names. Consequently, by a combination of circumstances the faunas of the various western Triassic beds were studied without much reference to each other, and so upon a basis of faunal lists the several assemblages appear to be rather different from each other.

There is good reason to think that the apparent dissimilarity of the faunas, as indicated by an uncritical listing of their component elements, is quite misleading. For example, a recent critical and biometrical study of the metoposaurid amphibians of North America presents evidence of the strongest kind to show that the 8 genera and 14 species of these

amphibians described from the Upper Triassic beds of this continent can be reduced to a single genus containing three species (Colbert and Imbrie, 1956). So far as amphibian evidence goes, this particular study emphasizes the general contemporaneity of beds of the Newark, Dockum, Chinle, and Popo Agie.

A similar study of the phytosaurs, now under way, indicates that nine genera of these reptiles which have been described can be reduced to four or five. The phytosaurs, however, offer more promise than do the amphibians of a basis for paleontological subdivision of the Late Triassic, as their skulls were undergoing rapid adaptive modifications. Morphologically primitive phytosaurs (*Paleorhinus* [including *Promystriosuchus*] and *Angistorhinus*) are confined to the Popo Agie of Wyoming and a few localities low in the Dockum of Texas. *Clepsysaurus*, *Heterodontosuchus*, *Machaeroprosopus*, *Leptosuchus*, *Pseudopalatus*, and *Brachysuchus*, all of which probably belong to a single genus of phytosaur, occur at higher levels in the Dockum Group, in the Chinle Formation, and in the Newark Group. Except for *Brachysuchus*, they are not known to occur with the more primitive phytosaur genera. Adaptive modifications of the skull make further subdivision of deposits containing phytosaurs possible. *Rutiodon*, known only from the Newark group, is intermediate in skull structure between *Paleorhinus* and *Phytosaurus*. Its stratigraphic significance cannot at present be more closely evaluated than "Upper Triassic."

Very likely similar conclusions might be drawn from critical studies of the other fossil vertebrates, hitherto described from the Upper Triassic beds of North America. Unfortunately the methods of analysis and synthesis that have been or are being applied to the fossil amphibians and to the phytosaurs cannot be applied to many of the other Triassic genera, since good population samples are not available.

Many apparent differences between the faunas under consideration are the result of absence of genera from all but one assemblage. These cases must be regarded with great caution, because probably differences of this type are mainly the result of accidents of collecting. If our samples were better, it seems quite likely that resemblances among the faunas would increase. This applies with particular force to the smaller elements in the faunas, which for the most part are known from scattered and fragmentary remains. Indeed, the picture that we have of these faunas at present is very incomplete and quite unbalanced; there is an abundance of two or three large forms, like the big amphibians and the phytosaurs, and a dearth of the smaller animals, which during Late Triassic time must have been rather numerous.

It may be useful to indicate some of the valid resemblances among these Late Triassic faunas of North America.

	Newark Group	Dockum Group	Chinle Formation	Popo Agie Member of the Chugwater Formation
Fishes				
Semionotus	x	....	x	....
Ceratodus	....	x	x	x
Amphibians				
Eupelor	x	x	x	x
Reptiles				
Desmatosuchus	....	x	x	....
Typothorax	....	x	x	....
Paleorhinus	....	x	....	x
Angistorhinus	....	x	....	x
Phytosaur*	x	x	x	....

* Studies now under way indicate that various North American Triassic phytosaurian genera may be reduced to synonymy with *Phytosaurus*.

These genera, as well as other genera found in the Newark, Chinle, Dockum, and Popo Agie, give clear evidence that these beds are the equivalents of the European Keuper. Therefore they may be considered as of Late Triassic age.

For example, *Semionotus* and *Ceratodus* are characteristic Keuper fishes in various parts of the world, although it must be admitted that the geologic range of both of these genera extends beyond the limits of the Upper Triassic. *Eupelor* is very closely related to *Metoposaurus* of the European Keuper. Some of the pseudosuchian reptiles show similar close relationships with genera in the Keuper of Europe. Thus *Stegomus* of the Newark may be compared with *Aetosaurus* of the Keuper, and a similar comparison is valid for *Hesperosuchus* of the Chinle and *Ornithosuchus* of the New Red Sandstone of Scotland. The phytosaurs of North America show very close comparisons with the old world phytosaurs of Keuper age. The same is true for some of the dinosaurs. Consequently the general age relationships of

the North American horizons that are here under consideration are quite clear.

Eastern North America.—The Newark Group, probably originally deposited in several separate basins and exposed in long belts with general northward trends, was studied long before any of the Triassic deposits of western North America had been discovered. Fossils were collected and described from the Newark more than a century ago; consequently comparisons and problems of synonymy in the study of Late Triassic vertebrates in North America must eventually rest in part upon those of the Newark Group.

In New Jersey and eastern Pennsylvania the Newark Group has long been subdivided into three divisions, in a general way from bottom to top, the Stockton, Lockatong, and Brunswick Formations. It was originally thought that these units of the Newark represented a time sequence, but in recent years there has been a growing tendency to regard these formations as indicating not an exact sequence in time, but rather a series of interfingering facies, in part contemporaneous with each other and only in part sequential. Thus, according to studies by McLaughlin, the Stockton is in part contemporaneous with the lower portion of the Brunswick. The Lockatong is entirely contemporaneous with a part of the Brunswick; it is a lacustrine facies within the torrential river and stream facies of the Brunswick.

In northern New Jersey the intrusive diabase that forms the Palisades along the Hudson River is included within the Stockton Formation. Beneath the diabase, in sandstones at about the level of the water, a partial skeleton of a phytosaur was found, early in the century. This was the specimen named *Rutiodon manhattanensis* by Von Huene, but is almost certainly referable to the genus *Clepsysaurus* (Colbert and Chaffee, 1941), a form closely related to the phytosaur genus of western North America, variously called *Machaeroprosopus*, *Leptosuchus*, etc. The coelacanth fish, *Diplurus*, has also been found in the Stockton in this general region, especially in a quarry at North Bergen (Schaeffer, 1952).

The Stockton Formation is succeeded by the Brunswick in the northern valley of New Jersey; the Lockatong is absent in this area. The Brunswick Formation is a thick sequence of sandstones and shales, and in its upper portion are several extrusive basalts that form prominent cliffs. The oldest of these extrusives is the First Watchung Mountain, a long ridge along the western side of the northern valley. Behind it is the Second Watchung Mountain, and behind that Hook Mountain. To the west the valley is bounded by a long fault. Fossil land-living vertebrates are indeed rare in the Triassic of New Jersey, as is generally the case in the eastern Triassic, so it is a rather unusual circumstance that two skeletons of the procolophonid reptile, *Hypsognathus* (Colbert, 1946), have been found at two localities only a few miles apart, and not far to the east of the First Watchung. Thus these two skeletons are stratigraphically beneath the first extrusive. At Boonton, New Jersey, west of and stratigraphically above Hook Mountain, the last of the extrusives, is the locality from which numerous specimens of subholostean and holostean fishes were collected, many years ago. This locality, long since under water as the result of the construction of a large dam, also yielded several specimens of *Diplurus*, the coelacanth fish mentioned as having been found in the Stockton.

Within recent years an enormously large series of *Diplurus* was uncovered during excavation for the new library building at Princeton University. These specimens were found in the Lockatong Formation, the northernmost limits of which are not far north of Princeton. The Lockatong, which extends southwesterly into Pennsylvania as a facies intercalated within the Brunswick, also yielded, some years ago, a lower jaw and a partial skull of *Clepsysaurus*, the phytosaur genus also found beneath the Palisades of the Hudson River in the Stockton Formation.

It is therefore evident that *Diplurus* and *Clepsysaurus*, at least, are long-ranging genera within the Newark Group, as it is exposed in New Jersey and Pennsylvania. The same is true for *Semionotus*, which has been found in the Stockton, beneath the Palisade diabase, and also at Boonton. It is probable that, if the fossil record in the Newark of New Jersey and Pennsylvania were more extensively known, the same considerations would apply to several if not most of the other Newark vertebrates.

One vertebrate of the Newark Group of particular significance is the metoposaur amphibian, *Eupelor*, from the Brunswick Formation at Phoenixville, Pennsylvania. As indicated, there is good evidence to show that all the metoposaurs of North America may be referred to this one genus, a fact that points to the close relationships between the sediments of the Newark and the Upper Triassic continental beds of the western states.

The Newark Group is exposed in an elongate belt in the Connecticut Valley from New Haven to northern Massachusetts. It is bounded on the east by a fault of large displacement, comparable to that which bounds the northern New Jersey area on the west. Further similarities between the two basins are seen in the dominantly red sediments, in the presence of basic intrusive rocks in the lower part, and in the three basic lava flows interbedded with the upper sediments. Dominantly arkosic red sandstones below the first lava sheet are called New Haven Arkose. The lavas and interbedded red and dark sediments constitute the Meriden Formation. Above the upper lava lies the Portland Arkose, consisting mainly of red sandstone and conglomerate. It seems quite logical to equate the three lava flows of the Meriden formation with the First Watchung, Second Watchung, and Hook Mountain of New Jersey. The West Rock Trap may be equivalent to the Palisades sill along the Hudson River.

Only fragmentary remains of a phytosaur and of a pseudosuchian reptile, *Stegomus*, have been found in the New Haven Arkose (Lull, 1953). Both genera also occur below the lavas in the Brunswick Formation of New Jersey, and *Clepsysaurus* occurs at even lower levels in the Stockton.

Black shales of the Meriden Formation have yielded the fossil fishes *Semionotus*, *Redfieldia*, *Dictyopyge*, *Ptycholepis*, and *Diplurus* (Eastman, 1911), all of which occur either in the Lockatong facies of New Jersey and Pennsylvania or in black shales immediately above the uppermost (Hook Mountain) lava flow at Boonton, New Jersey. Most of these genera also occur in dark shale beds of the Portland Arkose near Durham, Connecticut, and Chicopee Falls, Massachusetts. It has already been pointed out that *Diplurus* and *Semionotus* have wide stratigraphic range within the Newark Group, and it seems most reasonable to attribute their presence in the Meriden Formation of Connecticut and Massachusetts to appropriate facies rather than close time equivalence with fish-bearing levels of the New Jersey area. Reddish sandstones and shales within the Meriden Formation contain the first of the incredibly numerous dinosaur tracks that are so famous in the Connecticut Valley (Lull, 1953). Similar tracks are also known from the upper part of the Brunswick Formation at Boonton and Milford, New Jersey, but it should be emphasized that nowhere in the Connecticut Valley area have tracks of *Chirotherium* been found, though they are known from the Lockatong and horizons low in the Brunswick in New Jersey.

The Portland Arkose has yielded numerous tracks and four skeletons of dinosaurs (*Podokesaurus*, *Anchisaurus*) and a small pseudosuchian, *Stegomosuchus*. Fishes have been found where the black shale facies occurs.

The few occurrences of vertebrate fossils in the Newark Group in the Connecticut Valley and New Jersey areas are insufficient to demonstrate close correlation between individual formations in these areas. Their distribution suggests an earlier fauna of phytosaurs and pseudosuchians which was replaced at about the level of the lava flows (Meriden Formation, upper part of the Brunswick Formation) by a fauna dominated by bipedal dinosaurs. Apparent differences between the faunas of New Jersey and Pennsylvania and of the Connecticut Valley may be attributed largely to the accidents of preservation and discovery. Dinosaur skeletons are known from the Connecticut Valley but not from New Jersey. The amphibian, *Eupelor*, is known from Pennsylvania but not from the Connecticut Valley. Perhaps these differences are at least in part the result of facies differences.

Certain facies differences are important in comparing the vertebrate-bearing sediments of North Carolina with those farther north. In the southern region the vertebrates have been discovered in carbonaceous deposits of the Cumnock Formation, generally in coal mines (Emmons, 1856). *Dictyocephalus* is a small amphibian that at present is unique. Because of this it cannot be compared readily with the large metoposaur, *Eupelor*, from Pennsylvania. *Dromatherium* and *Microconodon* are mammal-like reptiles at present known from the two type jaws. Here again comparisons with elements of other Late Triassic faunas in North America are not possible. The one definable difference between the North Carolina fauna and the Newark fauna of more northerly regions is in the phytosaurs. *Rutiodon* is the phytosaur characteristic of North Carolina, and it is a type distinct from *Clepsysaurus* of New Jersey, Pennsylvania, and the Connecticut Valley. Moreover, *Rutiodon* is distinct from any of the Late Triassic phytosaurs of western North America.

The vertebrates of the Newark group are shown in Table 1.

Wind River Mountains, Wyoming.—Along the eastern flanks of the Wind River Mountains the Triassic exposures, generally designated in

TABLE 1.—VERTEBRATES OF THE NEWARK GROUP

Connecticut Valley		New Jersey-Pennsylvania	North Carolina
Portland Arkose	*Stegomosuchus* *Podokesaurus* *Anchisaurus* *Ammosaurus* Dinosaur tracks Fishes	*Semionotus* *Redfieldia* *Diplurus* *Clepsysaurus* Dinosaur tracks	Sanford Formation
Meriden Formation	*Dictyopyge* *Redfieldia* *Ptycholepis* *Semionotus* *Diplurus* Dinosaur tracks		
New Haven Arkose	*Stegomus* *Clepsysaurus*	Brunswick Formation *Hypsognathus* *Stegomus* *Eupelor* *Chirotherium* Dinosaur tracks	
		Lockatong Formation *Carinacanthus* *Turseodus* *Dictyopyge* *Redfieldia* *Semionotus* *Diplurus* *Eupelor* *Clepsysaurus* *Lysorocephalus* *Gwyneddosaurus* *Rhabdopelix*	Cumnock Formation *Diplurus* *Semionotus* *Dictyocephalus* *Rutiodon* *Dromatherium* *Microconodon*
		Stockton Formation *Turseodus* *Semionotus* *Diplurus* *Clepsysaurus*	Pekin Formation

their entirety as the Chugwater Formation, show a distinct division into three zones. The lowest division, the Red Peak Member, is closely comparable in its physical expression to the Moenkopi Formation of Arizona. Like the Moenkopi, the Red Peak consists of dark-red sandstones and shales, with frequent mud cracks and ripple marks, and some footprints. The Red Peak has not yielded bones of fossil vertebrates, a condition that was also typical of the Moenkopi until a comparatively recent date. Within the past decade or so parties from the University of California have uncovered fossil vertebrates at various localities within the Moenkopi, and it may well be that the bones of fossil vertebrates also will be found in the Red Peak. Because of these resemblances, it is reasonable to think of the Red Peak Member as approximately correlative with the Moenkopi Formation of Arizona.

In many regions the Red Peak Member of the Chugwater Formation is separated from the overlying Popo Agie Member by a comparatively thin but persistent limestone, the Alcova Limestone, and a sandstone, the Crow Mountain Sandstone. The limestone, containing scattered remains of *Corosaurus* (Case, 1936), a nothosaur, indicates a brief marine invasion of the Wyoming region and adjacent regions.

Finally, at the top of the Chugwater Formation, is the Popo Agie Member, soft reddish or yellowish siltstones with intercalated sandstones and conglomerates. In these sediments are found the bones of characteristic Upper Triassic vertebrates. The two commonest groups, as is true of the Upper Triassic in the southwest, are metoposaur amphibians (Branson and Mehl, 1929) and phytosaurs (Mehl, 1915b; 1928). The amphibians can all be included within the single genus *Eupelor*, the same genus that is found in the Chinle Formation and in the Dockum Group. This might point to a general identity of ages between the Upper Triassic sediments of Wyoming and those of the southwestern states. The phytosaurs of the Popo Agie, however, belong to the two genera, *Paleorhinus* and *Angistorhinus*, forms which are found in the lower part of the Dockum but not in the Chinle Formation, and which appear to be more primitive in certain characters than "*Machaeroprosopus*," the characteristic phytosaur of the Chinle Formation. Because of this, there may be some justification for supposing that the Popo Agie is of early Keuper age, comparable in this respect to the lower portion of the Dockum and definitely earlier than the typical Chinle of Arizona and New Mexico.

The remainder of the Popo Agie fauna is known from fragmentary fossils, so that comparisons with other Late Triassic faunas are difficult. *Dolichobrachium* is probably a pseudosuchian (Von Huene, 1939), but whether it is to be equated with the pseudosuchians of the Chinle and the Dockum is undetermined. *Brachybrachium* and *Eubrachiosaurus* appear to be therapsids (Von Huene, 1939), possibly related to *Placerias* of the Chinle formation. *Poposaurus* (Mehl, 1915b) is without doubt an ornithischian dinosaur, and as such it represents our best-known record of a Triassic ornithischian. What is of particular interest is the fact that *Poposaurus* shows some advanced ornithischian characters. Would this be evidence for placing the Popo Agie as a high rather than as a low part of the Upper Triassic sequence in North America? This question probably cannot be answered satisfactorily at present.

Northern Arizona.—In northern Arizona the Upper Triassic sedimentary rocks consist of an extensive sequence of sandstones, variegated shales, and limestones. Sedimentation probably was continuous here to the very end of Triassic time. Indeed, the Triassic of this area seems to merge into the Jurassic, so that it is difficult to draw a distinct line of demarcation between the sediments of the two periods. For example, the genus *Protosuchus* is found in the Dinosaur Canyon beds, now regarded as of Triassic age, but this genus is not in itself definitive, and it might very well be of Jurassic relationships. Above this, tritylodonts have recently been discovered in the overlying Kayenta Formation, now commonly regarded as of Jurassic age. Yet previously the tritylodonts have been found mainly in deposits of Late Triassic age.

The Late Triassic fauna of northern Arizona is, for all practical purposes, restricted to the Petrified Forest Member of the Chinle Formation. Among the most abundant fossils are the remains of *Eupelor*, closely comparable to *Eupelor* as it occurs in the Dockum Group, and a little less closely to *Eupelor* from the Popo Agie Member of the Chugwater Formation of Wyoming. The phytosaurs (Camp, 1930), also abundant in the Chinle as in other Upper Triassic beds of the west and the southwest, are represented by the genus "*Machaeroprosopus*," apparently a more advanced type than *Paleorhinus* and *Angistorhinus* of the lower portion of the Dockum and of the Popo Agie. There are three pseudosuchians in the Petrified Forest Member—*Hesperosuchus* (Colbert, 1952), a generalized pseudosuchian that may be compared with the Late Triassic genus, *Ornithosuchus*, of Europe, and *Desmatosuchus* and *Typothorax*, large, specialized types, found also in New Mexico and in Texas. *Placerias* is a dicynodont therapsid.

Unfortunately, the vertebrates of the Petrified Forest Member other than the phytosaurs give little definite evidence as to the position within the Upper Triassic to which this zone should be assigned. It is certainly of Upper Triassic or Keuper relationships. It may be in part somewhat later than the Popo Agie and the lower portion of the Dockum sequence. It may be equivalent to a part of the upper section of the Dockum. It is almost certainly not the latest part of the Triassic in North

TABLE 2.—VERTEBRATES OF THE UPPER TRIASSIC OF WYOMING, ARIZONA, AND NORTHERN NEW MEXICO

Wind River Mountains, Wyoming			Northern Arizona			Northern and central New Mexico		
			Moenave Form.	Springdale Member	*Semionotus*			
				Dinosaur Canyon Member	*Protosuchus*			
			Wingate Sandst.	Lukachukai Member				
				Rock Point Member	Phytosaurs			
			Chinle Formation	Owl Rock Member	Phytosaurs			
				Petrified Forest Member	*Semionotus* *Ceratodus* *Eupelor* *Hesperosuchus* *Typothorax* *Desmatosuchus* *Machaeroprosopus* *Placerias*	Chinle Formation	Correo Sandstone Member	Metoposaurs Phytosaurs
							Petrified Forest Member	*Eupelor* *Typothorax* *Machaeroprosopus* *Coelophysis*
				Lower ss. Member			Poleo Ss. M.	
				Shinarump Member	Phytosaurs Metoposaur Fragments		Salitral Sh. Tongue Agua Zarca Ss. M.	
Chugwater Formation (part)	Popo Agie Member	*Ceratodus* *Eupelor* *Dolichobrachium* *Paleorhinus* *Angistorhinus* *Poposaurus* *Brachybrachium* *Eubrachiosaurus*						

America. Perhaps it may be thought of as representing approximately a middle segment of Upper Triassic or Keuper sedimentation on this content.

Northern New Mexico.—The Triassic of northern New Mexico is of less vertical extent than it is to the west in Arizona, or to the east in Texas. The Chinle Formation consists of a fairly thick sequence of variegated shales above some comparatively thin sandstones and shales; the latter are designated, from bottom to top, the Agua Zarca Sandstone Member, the Salitral shale Tongue, and the Poleo Sandstone Lentil. At the base of the Agua Zarca is an unconformity separating it from the underlying Permian rocks, and at the top of the Chinle in this region is another unconformity separating the Triassic from the overlying Jurassic rocks. To the south the Correo Sandstone Member is present in the top of the Chinle, beneath the upper unconformity. The Triassic fauna in this area is found at various levels within the Chinle, above the Poleo Sandstone.

The Chinle in New Mexico has yielded numerous remains of the amphibian, *Eupelor*, and "*Machaeroprosopus*," in which respect it resembles the Chinle of Arizona. The phytosaurs are similar to those in the higher fossiliferous levels of the Petrified Forest Member in Arizona. The large armored pseudosuchian, *Typothorax*, known from Arizona is represented in northern New Mexico by fairly complete skeletal material (Gregory, 1953).

Finally, the theropod dinosaur, *Coelophysis*, a form that occurs sparsely in the Chinle of Arizona, is known from near the top of the Chinle of New Mexico by abundant, completely articulated skeletons.

From these occurrences it seems apparent that the fossiliferous Chinle Formation of New Mexico is correlative with the upper part of the Petrified Forest Member of the Chinle in Arizona. It can therefore be presumed to have about the same age relationships as the exposures in Arizona, which means that it represents about the middle portion of the Upper Triassic, or Keuper.

Table 2 shows the vertebrates of Wyoming, Arizona, and northern New Mexico.

Eastern New Mexico.—The Dockum group of eastern New Mexico and western Texas consists of dominantly red siltstones and claystones containing minor sandy lenses and thick beds of cross-bedded light-colored sandstone and conglomerate which underlie the Staked Plains and southern High Plains and are exposed around their margins. They overlie various Permian red beds. No lower or Middle Triassic vertebrates have been found even in the lowest beds of the Dockum.

South of Lamy a quarry at of unknown depth in the Dockum has yielded numerous skeletons of the amphibian *Eupelor*.

In the vicinity of Tucumcari, New Mexico, the Santa Rosa Sandstone forms the base of the Triassic; it is exposed along the canyon of Canadian River. A thick sequence of red clays interrupted by some light-colored massive sandstones overlies the Santa Rosa. It has been called a member of the Chinle Formation, which seems an undesirable restriction of the term Chinle because the Dockum Group as a whole appears to be equivalent to the Chinle Formation of Arizona. The upper 400 feet of the Dockum in this area is called the Redonda Member of the "Chinle" and consists of persistent thin beds of fine orange-red sandstone alternating with variegated shales and limestone.

Fossil vertebrates have been obtained from the upper part of the red clay member of the "Chinle" Formation 80 to 200 feet below the base of the Redonda Member. An advanced species of phytosaur from this level is comparable with those from the Ghost Ranch locality in northwestern New Mexico and from the uppermost fossiliferous levels of the Petrified Forest Member of the Chinle at Adamana, Arizona. *Typothorax* is present, but *Eupelor* has not been found at this level. Most of the specimens have been found near rather persistent sandstones which are possibly correlative with the Trujillo Formation of Texas. Hence they are stratigraphically higher than the fossiliferous localities on Sierrita de la Cruize northwest of Amarillo.

Abundant phytosaur teeth and other fragments occur in a bone bed a few feet above the base of the Redonda member. A phytosaur skull from just below this bone bed is more specialized in its temporal region than any described species.

Along the Cimarron River in northeastern New Mexico the Sheep Pen Sandstone and Sloan Canyon Formation are recognized above the undifferentiated shales of the Dockum Group. These two upper units are lithologically similar to the Redonda, and a phytosaur (Stovall and Savage, 1939) from the Sloan Canyon appears identical with that from the Redonda.

TABLE 3.—VERTEBRATES OF EAST-CENTRAL NEW MEXICO, NORTHEASTERN NEW MEXICO, COLORADO-OKLAHOMA, AND TEXAS

Equivalents			East-central New Mexico			Northeastern New Mexico and Colorado-Oklahoma		Texas		
Moenave Fm.	Springdale Sandstone Member									
	Dinosaur Canyon Member									
Wingate Ss.	Lukachukai Member									
	Rock Point Member				Redonda Member	Phytosaur[a] undescribed	Sheep Pen Ss.			
							Sloan Canyon Form.	Phytosaur[a]		
Chinle Formation	Owl Rock Member	*Machaeroprosopus*[b]		"Chinle" Formation	Shale Member	*Machaeroprosopus* *Brachysuchus*[b] *Typothorax*	Shale unit		Trujillo Form.	*Leptosuchus*[c] *Typothorax* *Eupelor* Coelacanth
	Petrified Forest Member	*Machaeroprosopus*[c]								
	Lower sandst. Member		Dockum Group		Santa Rosa Sandstone			Dockum Group		
	Shinarump Member	*Paleorhinus*[d] *Angistorhinus*							Tecovas Form.	*Paleorhinus* *Angistorhinus* *Brachysuchus* *Typothorax* †*Desmatosuchus* *Trilophosaurus* Saurischian dinosaur *Eupelor* *Latiscopus* *Ceratodus* Catopterid *Colognathus*
	Popo Agie Member (Chugwater Formation)									

[a] Phytosaur with completely concealed superior fenestra.
[b] *Machaeroprosopus tenuis* Camp and *M. andersoni* (Mehl).
[c] *Machaeroprosopus adamanensis* (Camp), *P. doughtyi* Case, *L. crosbiensis* Case, etc.
[d] *Paleorhinus* fauna. Primitive *Phytosaurus* (*Brachysuchus*) only.

Texas.—In the Canadian River Valley the Dockum Group has been divided into the Tecovas Formation below and Trujillo Formation above. The Trujillo has sometimes been correlated with the Santa Rosa Sandstone, but there is not proof of this and it seems more probable that the Santa Rosa either thins eastward and disappears before reaching the Texas line or else interfingers with the silts and shales of the Tecovas, which rest directly on the Permian.

An important quarry in the type Tecovas shale on Sierrita de la Cruize Creek northwest of Amarillo has produced a fauna comparable to that of lower levels of the Petrified Forest Member of the Chinle Formation in the

Petrified Forest region. *Eupelor*, "*Leptosuchus*," *Typothorax*, and "dinosaurs" occur here (Case, 1932; Case and White, 1934; Colbert and Imbrie, 1956). Both *Eupelor* and the phytosaur are indistinguishable from the Chinle species.

Teeth and other fragments of phytosaurs, and a coelacanth skull are known from the Tecovas Formation in Palo Duro Canyon.

Farther south along the east side of the Staked Plains the Tecovas Formation appears to thin below the "medial sandstones," which have been correlated with both the Trujillo and Santa Rosa. In Crosby and Dickens counties a thin basal conglomerate appears, the northern end of a member better developed to the South.

Numerous vertebrate fossils have been recovered from the Dickens-Crosby County area, and these appear to represent two faunas, one similar to that of Sierrita de la Cruize, the other, more primitive, known from localities farther south in the lower portion of the Dockum Group. Unfortunately no records are available of the exact sites from which different specimens were obtained in Crosby County, so the stratigraphic relations of these faunas, if they are indeed distinct, are not yet known. The fauna from Crosby County (Case, 1922) consists of: Coelacanth fish, *Colognathus*, *Ceratodus*, *Eupelor*, *Typothorax*, *Desmatosuchus*, "*Leptosuchus*", "*Brachysuchus*", *Paleorhinus*, *Trilophosaurus*.

In southeastern Howard County (Gregory, 1945; Sawin, 1945) exposures several hundred feet above the base of the Dockum Group have yielded a fauna similar to that of the Popo Agie Member of the Chugwater Formation of Wyoming. Many of these fossils also are found in Crosby County. In Scurry County, near the base of the Dockum, fragments of the primitive phytosaur *Paleorhinus* (Langston, 1949) were obtained, and presumably at a higher level a group of the amphibian, *Eupelor*, were found. On the north slope of the Glass Mountains the Bissett Conglomerate has yielded a horn of the reptile *Desmatosuchus*.

The fauna from Howard County includes: "*Brachysuchus*" ,*Angistorhinus*, *Paleorhinus*, *Typothorax*, *Trilophosaurus*, *Eupelor*, *Latiscopus*, Catopterids.

Table 3 shows the vertebrates from east-central New Mexico, northeastern New Mexico, and Texas.

Rhaetic Deposits

It is here assumed that the Rhaetic constitutes an uppermost division of the Triassic, above the Keuper. If such be the case, several continental formations in the southwestern United States may be assigned to the Rhaetic.

The Wingate Sandstone, as now recognized by the U. S. Geological Survey (Averitt *et al.*, 1955), the overlying Moenave Formation, and the Kayenta Formation, on the basis of extremely meager remains of vertebrates, seem equivalent to the German Rhaetic.

The Wingate as designated by the United States Geological Survey in northeastern Arizona has been subdivided into two members, the Rock Point below and the Lukachukai above. The Rock Point, which can be equated with the "Chinle A" of H. E. Gregory, has within recent years yielded unmistakable phytosaur teeth. Consequently there can be no doubt of the Triassic age of this zone.

The Moenave Formation likewise has been subdivided into two members, the Dinosaur Canyon below and the Springdale Sandstone above. In the Dinosaur Canyon fossils of a primitive crocodilian, *Protosuchus*, have been discovered (Colbert and Mook, 1951.) The age of this fossil can be interpreted either as Triassic or Jurassic, according to the bias of the student. It is closely related to primitive crocodilians in South Africa which are commonly regarded as of uppermost Triassic age.

The characteristic Triassic fish, *Semionotus*, has been found at the top of the Dinosaur Canyon Member and in the Springdale Sandstone Member at several localities in northern Arizona and south Utah (Hesse, 1935; Schaeffer and Dunkle, 1950).

The age of the Kayenta Formation constitutes a baffling problem. Recently Welles (1954) described a dinosaur from the lower portion of the Kayenta Formation, and he referred this new specimen to the Jurassic genus, *Megalosaurus*. Whether this dinosaur, which is known from a most excellent skeleton, is to be assigned properly to the genus *Megalosaurus* is a debatable question, a point recently raised by Swinton (1955), who states that "its relationships are certainly with the early carnosaurs, but it is not *Megalosaurus*." However that may be, the dinosaur from the Kayenta has a definite "Jurassic look," as indicated by its deep, carnosaur-type skull, its reduced forelimbs, and by certain other anatomical characters.

Against this evidence are the recently discovered tritylodont skeletons, found within about 10 feet of the top of the Kayenta Formation. G. E. Lewis, who first excavated these very important fossils, has indicated that they are most closely related to the tritylodont *Bienotherium*, of the Lufeng Series of China, and to *Tritylodon*, of the Stormberg series of South Africa (*in* Averitt *et al.*, 1955, p. 2523). The Lufeng is generally regarded as of Keuper age, as is the Stormberg, although some authors would include the latter in the Rhaetic or even in the Liassic. Certainly the arguments for regarding the new tritylodonts from the Kayenta as of Triassic affinities are very strong indeed, yet these fossils come from a level above that at which the "*Megalosaurus*" skeleton was found.

Thus the various lines of evidence as to the age of the Kayenta are conflicting. If we base our conclusion upon the newly discovered tritylodonts, then we are led to the view that the Kayenta should probably be included within the Rhaetic, above the Moenave. On the other hand, the dinosaur skeleton described by Wells might lead us to think that the Kayenta is of Jurassic age, so that the dividing line between this formation and the underlying Moenave might represent the boundary between Triassic and Jurassic in the Plateau region. There the matter rests. Perhaps additional fossil evidence may help to clarify this problem.

ANNOTATIONS

1. *Column 1. Nicaragua and Honduras.*—Willis (1912, p. 500–502) and Schuchert (1935, p. 356, 362) reviewed the nonmarine deposits in Nicaragua and Honduras and attributed them to the Rhaetian on the basis of fossil plants identified by Newberry (1888a; 1888b). Most of the record of these deposits is due to Sapper (1905, p. 52) and Mierisch (1895, p. 61, 62). Sapper expressed some doubt about the age assignment, but Willis and Schuchert accepted it. Knowlton (1918) considered the age to be either Late Triassic or Jurassic. Maldonado-Koerdell (1953) described a Rhaetic-Liassic plant and other fossils from the same formation, to be correlated with similar beds in southern and central Mexico. Assemblages of fern and cycad remains such as Newberry identified here, are widely distributed in Central America, Mexico, and the United States, but their age is not well determined. In the present state of knowledge it is not possible to state precisely what level is represented in the deposits, and it seems better to show them in this chart as possibly including Upper Triassic beds. (*Maldonado-Koerdell*)

2. *Column 2. Chiapas.*—The Todos Santos Formation was first described by Dollfus and Mont-Serrat (1868) and was later named in northwestern Guatemala (Sapper, 1896), but appears to be widely distributed in Chiapas also. It contains gray, red, and yellow conglomerates, sandstones, and shales, and is nearly barren of fossils, though the upper part has yielded cycad remains that associate it with the widespread flora commonly assigned to the Rhaetian. Opinions of the age of the formation have varied greatly; Aguilera (1907) included it in his Neotriassic Series, Ver Wiebe (1925) placed it partly in the Triassic and partly in the Jurassic, Mullerried (1942) placed it in the Jurassic, and Schuchert (1935, p. 329) called it Lower Cretaceous and possibly older, attributing this opinion to Boese (1905). (*Maldonado-Koerdell*)

3. *Column 3. Oaxaca and Guerrero.*—Salas (1949, p. 91–95) described the Mixtepec-El Consuelo beds—arenaceous shales and sandstones with intercalations of carbonaceous materials contining numerous fragments of ferns and cycads. These rest upon crystalline rocks or on shales of the "basal section" and are overlain by Tertiary shales or Cretaceous limestones. These beds had been recorded by various previous writers. (*Maldonado-Koerdell*)

4. *Column 3. Oaxaca and Guerrero.*—The Cualac Quartzite was described by Guzmán (1950, p. 108–112) as a nearly pure conglomeratic quartzite, varying much in thickness, absent at some places and reaching a maximum thickness of 400m (1300 feet). It rests on a crystalline complex and is overlain by the beds with fossil ferns and cycads. It is without fossils but has been assigned to the Triassic on the basis of field relations with overlying rocks of younger age. (*Maldonado-Koerdell*)

5. *Column 3. Oaxaca and Guerrero.*—The basal section of clastics and clays was described by Salas (1949, p. 85–92) as composed at places of a basal conglomerate, pyroclastic breccias, coarse sandstones, and conglomeratic sandstones, red to gray and occasionally green; and at other places of hard shales, in part pure, in part sandy, gray, green, or brown. It rests on crystalline rocks and is overlain by the de-

posits with fossil ferns and cycads. It is without fossils. (*Maldonado-Koerdell*)

6. *Column 4. Puebla.*—Aguilera (1896, p. 90) described the rocks of the Matzitzi Formation as including conglomerates, sandstones, and shales, containing plant fossils and beds of coal. In 1907 he placed them in his Neotriassic Series. Burckhardt (1930) assigned at least the upper part of the beds with plants in southern Puebla and adjacent parts of Oaxaca and Guerrero to the middle and upper Liassic, leaving the remainder unassigned.

(*Maldonado-Koerdell*)

7. *Column 5. Huasteca.*—The name Huayacocotla Formation was proposed by Imlay and others (1948, p. 1750), with a type locality in northeastern Puebla. It is dominantly dark shale with some intercalated sandstone and conglomerate and a few lenses of limestone. The upper part is both marine and nonmarine and of Liassic age, and the lower part is plant-bearing and nonmarine. The lower limits of the formation are not known, and it is assumed that it rests on metamorphic rocks. Various writers have discussed the formation, including Bárcena (1875), Aguilera (1896), Díaz Lozano (1916), Burckhardt (1930), and Imlay (1943).

(*Maldonado-Koerdell*)

8. *Column 6. Zacatecas.*—Maldonado-Koerdell (1948, p. 291–293) summarized the work of Burckhardt (1905; 1906a; 1906b; 1930) and of Gutiérrez Amador (1908) on the Triassic rocks in Zacatecas; he also described some new fossils. The beds with Carnian marine fossils are blue or black siliceous and argillaceous shales alternating with green sandstones and green clays, which appear to be tuffaceous in part, and higher beds of quartzite and dark-gray quartzitic sandstones. The fossiliferous strata are overlain by greenstones of submarine volcanic origin interbedded with marine sediments. Burckhardt considered these part of the Triassic System and interpreted another later unconformable series of greenstones to be also part of the Triassic. The Triassic rocks rest on an undated sequence of steel-gray sericite schists and are succeeded by a pronounced hiatus. (*Maldonado-Koerdell*)

9. *Column 7. Chihuahua.*—In an unpublished report by geologists of Petróleos Mexicanos, Maldonado-Koerdell, for an area located southeast of Chihuahua City, ascribed to the Upper Triassic a thick series of black siliceous and argillaceous shales, alternating with green sandstones and green clays, containing internal molds of pelecypods and overlain by Jurassic fissile shales and Cretaceous rudistid limestones. Correlation with the Triassic beds of Zacatecas seems indicated on the basis of lithologic and paleontologic evidence.

(*Maldonado-Koerdell*)

10. *Column 8. Central Sonora.*—The name Barranca was given by Dumble (1900, p. 139–143) to a unit in central Sonora, a sequence of sandstones, quartzitic sandstones, conglomerates, and beds of shale that contain abundant remains of plants, as well as beds of coal and graphite. He assigned it to the Triassic. Aguilera (1896) had previously referred the deposits to the Upper Triassic, and Newberry, as early as 1876, had assigned plants from this area to the Rhaetian. R. E. King (1939, p. 1645, 1651–1658) redescribed the formation and extended it from the Triassic to the Lower Jurassic.

(*Maldonado-Koerdell*)

11. *Column 8. Central Sonora.*—The sandstones and shales with coal and fossil plants found in south-central Sonora are probably local expressions of the Barranca Formation. *See* Annotation 10. (*Maldonado-Koerdell*)

12. *Column 9. El Antimonio, Sonora.*—The section near El Antimonio has been described by Burckhardt (1930, p. 6) and Schuchert (1935, p. 144), attributing the data to W. T. Keller and C. L. Baker, respectively, and by White and Guiza (1949, p. 88–89). Ammonites of Liassic age were collected by White and Guiza about 6 km southeast of El Antimonio from the top of a sequence of beds containing Upper Triassic marine fossils. The Liassic fossils were about 50 m (165 feet) above a conglomerate of quartzite and quartz porphyry containing cobbles up to 15 cm (6 inches) in diameter, about 250 m (800 feet) above the highest Triassic fossils noted and about 500 m (1600 feet) above a horizon of Carnian fossils listed by Burckhardt. No thickness is given by Burckhardt for his unit no. 5, containing the Norian fossils listed by him, but the suggestion from the section of White and Guiza is strong that it is about 250 m (820 feet), with a discrepancy in the two sections of about 425 m (1400 feet). The highest Triassic fossils collected by White and Guiza are no younger than Noric, and no evidence is known of a Rhaetian flora. Further study of the area is needed. (*Maldonado-Koerdell*)

13. *Column 9. El Antimonio, Sonora.*—Burckhardt (1930, p. 6) and Schuchert (1935, p. 144) assign a middle part of the sequence to the Norian, principally on the basis of recognition of *Monotis* (as *Pseudomonotis*) *sub-*

circularis Gabb. Keller (1928) mentions the occurrence of sandstone, schist, sandy limestone, and massive limestone in this area, with Norian fossils (*Myophoria, Avicula, Lima*) in addition to *M. subcircularis*. The thickness of the interval assigned is indeterminable, but the complete section given by White and Guiza (1949, p. 88–89) suggests that the thickness would not be greatly different from 450 m (1500 feet). The highest fossils collected by White and Guiza do not include diagnostic forms (*fide* J. B. Reeside, Jr.) but do not in any way suggest a Rhaetian age. It seems appropriate to regard the sequence to the top of the conglomerate as of Norian age, in which case the thickness given by White and Guiza would be about 475 m (1550 feet) and that by Burckhardt about 900 m (2900 feet). (*Maldonado-Koerdell*)

14. *Column 9. El Antimonio, Sonora.*—Keller (1928), Burckhardt (1930, p. 6), and Schuchert (1935, p. 144) assign the unit with Carnian fossils and the lower beds down to the top of the Permian to the Carnian, a total of about 200 m (650 feet). The corresponding thickness in the section by White and Guiza (1949, p. 88–89) is 1250 m (4100 feet), divided by them into four major units, the upper three of which have yielded marine fossils. The lowest unit contains some coarse conglomerate and rests upon Permian beds. (*Maldonado-Koerdell*)

15. *Column 10. Santa Ana Mountains, California.*—The name Bedford Canyon Formation was proposed by E. S. Larsen (1948, p. 18) for a "group of mildly metamorphosed slates and argillites with some quartzites and a very few thin lenses of limestone" in the Santa Ana Mountains, with an estimated exposed thickness of 20,000 feet. The presence of abundant brachiopods of the genus *Halorella* established the Norian (Upper Triassic) age of at least a portion of this formation, although, as mentioned by Larsen, "some Jurassic rocks and some rocks older than the Triassic may be included."

Fossils collected by H. W. Fairbanks from a limestone in the Santa Ana Mountains were identified as "*Pseudomonotis* aff. *P. clarai*" of Early Triassic age by Smith (1898, p. 779), who introduced the name "Santa Ana limestone" for this unit. Subsequently Smith (1914, p. 145) changed his identification and assigned these pelecypods to "*Daonella sanctae-anae* Smith, *sp. nov.*", supposedly of Middle Triassic age. This generic assignment seems doubtful, and irrefutable evidence for the presence of Middle or Lower Triassic rocks in the Santa Ana Mountains has not yet been produced.

Merrill (1914, p. 639) used the local name "Santa Ana" for the slates of the Santa Ana Mountains. However, since the name Santa Ana has been applied to Cenozoic rocks in California, the name Bedford Canyon Formation was substituted by Larsen.

Portions of the metamorphic rocks assigned to the "Julian Group" or "Julian Schist" in the Peninsular Ranges of San Diego County have been considered by Merrill (1914, p. 640) and others as possible equivalents of the rocks assigned to the Bedford Canyon Formation, although fossil evidence is lacking. Likewise, the Santa Monica slates of the Santa Monica Mountains in Los Angeles County have questionably been referred to the Triassic (Hoots, 1931, p. 88). (*Silberling*)

16. *Column 11. Eastern San Bernardino County, California.*—Hazzard (1937) has reported the occurrence of a small area of Lower Triassic marine sediments in the Providence Range of eastern San Bernardino County. Approximately 1000 feet of limestone and clastic sediments, from which "very late Lower Triassic" fossils were collected, lies unconformably on Permian limestone. This section is overlain by Cenozoic sedimentary and volcanic rocks, and no record is available of younger Triassic deposits which may have been present in the area. (*Silberling*)

17. *Column 12. Mineral King area, Tulare County, California.*—The metamorphic rocks which compose a roof pendant of the Sierra Nevada at Mineral King are described by Knopf and Thelen (1905, p. 242) as consisting "prevailingly of clay slates and phyllites with interbedded limestones, quartzites, and large masses of tuffs and sheared volcanics, both acid and basic." The age of these rocks was provisionally set as Triassic on the basis of fossils examined by T. W. Stanton (Turner, 1894a, p. 451). Fossils collected by Durrell and identified by S. W. Muller include the ammonite genera *Juvavites* and *Arcestes* and established the Late Triassic age for at least a portion of these rocks (Durrell, 1940, p. 17).

Metamorphosed volcanic and sedimentary rocks in the foothills of the Sierra Nevada about 20 miles to the west of Mineral King, assigned to the Kaweah Series by Durrell (1940, p. 13), are similar in lithology and possibly equivalent in age to the Triassic rocks of the Mineral King roof pendant. Further evidence for at

least the Mesozoic age of the Kaweah Series is the fact that in the Inyo Mountains, approximately 55 miles to the east, pliomagmatic sediments are restricted to deposits younger than Early Triassic.

Metamorphosed Triassic deposits are probably present in a number of the roof pendants of the southern Sierra Nevada; however, faunal evidence for their age has not yet been discovered. (*Silberling*)

18. *Column 13. Inyo Mountains, California.*—The Triassic section of the Inyo Mountains consists of fossiliferous marine sediments of Early and Middle Triassic age overlain by a complex of volcanic and clastic sedimentary rocks several thousand feet in thickness. These rocks have been briefly described by Smith (1904, p. 350–352) and Kirk (1918, p. 47–48), but no detailed description of their stratigraphy has been published. Ammonoids from two different horizons in the marine section of Early Triassic and early Middle Triassic age have been described by Hyatt and Smith (1905) and by Smith (1914; 1932).

The marine calcareous shales and limestones with a thickness of more than 1500 feet are overlain by several thousand feet of volcanic rocks with subordinate conglomerates, sandstones, shales, and mudstones, considered to be of terrestrial origin by Merriam (1955). No unconformity or evidence of erosion has been recognized between the marine sediments and the overlying volcanic sequence so that at least the lower portion of the volcanic rocks is thought to be of Middle Triassic age. The upper part of this section may include rocks of Late Triassic or even post-Triassic age.

The base of the marine Triassic section overlaps the Permian rocks of the Inyo Mountains and unconformably overlies sediments of both Permian and Pennsylvanian age (Merriam, 1955). (*Silberling*)

19. *Column 14. Northern Sierra Nevada, California.*—The Milton Formation and the Sailor Canyon Formation, originally described from exposures respectively in the Downieville quadrangle and in Placer County by Turner (1894b, p. 232-234), include a thick section of metamorphosed sedimentary rocks of Late Triassic and Jurassic age. Taliaferro (1942, p. 99–100) considers these units to be equivalent and applies the name Milton Formation to "all the Triassic and Jurassic rocks of the higher parts of the Sierra Nevada, south of Taylorsville." (*Silberling*)

20. *Column 15. Taylorsville area, Plumas County, California.*—The Hosselkus Limestone was first described in the Taylorsville region by Diller (1892, p. 374) as overlying the Swearinger Slate, in which the Late Triassic pelecypod *Monotis* was found. On the basis of a fauna which was at first thought to be equivalent in age with *Monotis*, Smith (1894, p. 604–609) adopted the name Swearinger in Shasta County for slates subsequently included in the Pit Formation. Because of the apparent agreement with the Taylorsville section, Smith applied the name Hosselkus to the superjacent limestone unit in Shasta County. Later, *Monotis* was discovered in the slates, for which the name Brock Shale was eventually proposed by Diller (1906); the shale overlies the so-called Hosselkus Limestone in Shasta County. Thus the unit referred to the Hosselkus in Shasta County was apparently older than strata containing *Monotis*, and the type Hosselkus was younger. In order to make the section containing the type Hosselkus Limestone in the Taylorsville area conform with the supposed formational sequence in Shasta County, Diller (1908, p. 33) decided that the Taylorsville section was overturned and that the Swearinger Slate was stratigraphically superjacent to the type Hosselkus Limestone. The fossils reported by Hyatt (1892, p. 400) from the Hosselkus Limestone at Taylorsville are indicative only of a Late Triassic age.

Recent work in the Taylorsville area by Vernon McMath (Personal communication, 1955) indicates that the sequence of the Swearinger and Hosselkus Formations is probably in an upright position as originally described by Diller. This viewpoint is based on sedimentary structures observed within the Swearinger slates. Likewise the Trail Formation is evidently upright and according to McMath appears to underlie the Swearinger gradationally. Triassic rocks younger than the Hosselkus are not exposed in the Taylorsville area, owing to faulting, and the contact between the Trail Formation and the Kettle Meta-andesite is tentatively regarded as a fault. Permission from Mr. McMath to present these findings is gratefully acknowledged.

The name Cedar Formation was used in the Lassen Peak folio by Diller (1895) for a unit consisting primarily of limestone and containing a fauna similar to that of the Hosselkus Limestone in the Taylorsville region. (*Silberling*)

21. *Column 16. Shasta County, California.*—Although the strata referred by Diller (1906)

and Smith (1927, p. 4) to the Pit Shale, Hosselkus Limestone, and Brock Shale in Shasta County are sufficiently fossiliferous to establish their age relations, the ages of the underlying and overlying formational units have been the subject of some controversy. The Dekkas Andesite was originally included in the Triassic by Diller (1906), but Wheeler (1939b, p. 371) placed this unit in the Permian after re-evaluating its stratigraphic relationships. This assignment was confirmed by Albers and Robertson (1952) who discovered Permian fossils in sedimentary lenses of the Dekkas Andesite.

Diller (1906) described the Bully Hill Rhyolite as underlying the Pit Formation with some interstratification of the two units. Subsequently some of the rocks assigned to the Bully Hill Rhyolite by Diller were considered to be intrusive by Graton (1910, p. 81), and the formational name was discarded as unnecessary. This interpretation was followed by several later authors. Recently Albers (1953, p. 5–7) has re-established the name Bully Hill Rhyolite and presented evidence for the extrusive origin of these rocks.

The name Hosselkus for the limestone unit exposed between Squaw Creek and the Pit River as originally applied by Smith (1894) may not be appropriate. As pointed out in the discussion in Annotation 20 of the column at Taylorsville, Plumas County, there is some evidence that the Hosselkus at its type locality represents a higher stratigraphic position than the so-called Hosselkus Limestone in Shasta County, which was treated as a correlative by Diller (1906) and Smith (1927). The lenticular nature of the "Hosselkus Limestone" in Shasta county was attributed by Diller (1906) to an interlensing of shale and limestone facies of essentially contemporaneous rocks. Sanborn (1952) has suggested that the Brock Shale may disconformably overlie the "Hosselkus Limestone" in Shasta County, in which case the discontinuous nature of the "Hosselkus" may be due to a short period of erosion immediately following its deposition.

The Modin Formation was considered by Diller (1906) to overlie unconformably the Brock Shale with only a small degree of overlap and was assigned to the Lower Jurassic although definite paleontologic evidence for its age was not available. Sanborn (1952) has studied this unit in some detail and reports Norian (Upper Triassic) fossils from a zone about one-third of the distance above the base of the section and the characteristic Rhaetian ammonite *Choristoceras marshi* from the uppermost part of the section. According to Sanborn the total thickness of the Modin Formation, composed of argillaceous and volcanic rocks, conglomerates, and some massive limestones, is approximately 5500 feet. Overlying the Modin Formation are volcanic and clastic rocks which contain an Early Jurassic (Sinemurian) fauna at their base, but strata representing the base of the Jurassic appear to be absent. Consequently, the Triassic-Jurassic boundary in this area is considered to be an unconformity (Sanborn, 1952). (*Silberling*)

22. *Column 17. Clark County, Nevada.*—The Triassic rocks of Clark County in southern Nevada have been assigned to the Moenkopi, Shinarump, and Chinle Formations by Longwell (1928) in the Muddy Mountains and by Hewett (1931) in the Goodsprings quadrangle. The sections from each of these two localities appear to be closely related. The formational designations applied are derived from the Plateau region to the east, and in each case the Lower Triassic Moenkopi Formation unconformably overlies the Kaibab Limestone of Permian age, and the Upper Triassic Chinle Formation is overlain by Jurassic (?) rocks assigned to the Aztec Sandstone by Hewett (1931, p. 35). No definite information is available on the age of the Shinarump Conglomerate or the upper limit of the Moenkopi Formation in the area. S. W. Muller (Personal communication) is of the opinion that the marine beds in the upper part of the Moenkopi may be of Middle Triassic age. (*Silberling*)

23. *Column 18. Hawthorne and Tonopah quadrangles, Nevada.*—The Hawthorne and Tonopah quadrangles occupy portions of Mineral, Nye, and Esmeralda Counties in western Nevada and extend into California in the vicinity of Mono Lake. The Triassic System in this area has been described in some detail by Muller and Ferguson (1936; 1939), and the age relationships of some of the formational units are fairly well established.

The Upper Triassic Luning Formation, the base of which transgresses time planes to some extent across the area, overlies the dominantly volcanic Excelsior Formation with angular unconformity in the central part of the area. To the northeast, however, the Luning rests with erosional disconformity upon sediments of the Grantsville Formation, which in turn overlies volcanic rocks of Permian age. Recent collections by the writer indicate a late Middle

Triassic age for the Grantsville Formation. The Excelsior Formation has been assigned a probable early Middle Triassic age by Muller and Ferguson (1939, p. 1589) on the basis of fossils from a single locality in the northern Gillis Range. The so-called Excelsior at this locality is not in depositional contact with any of the other pre-Tertiary formational units recognized in the area, and possibly it has been introduced into the area by thrusting. If so, other exposures of the Excelsior, including those at the type area in the Excelsior Mountains, may be equivalent to the Permian volcanic rocks outcropping farther east.

In the southern part of the area the Candelaria Formation of early Early Triassic age overlies sedimentary rocks of Permian age. These rocks are nowhere in contact with the Excelsior-Luning sequence, and presumably they are on a different plate of a major thrust fault (Ferguson, Muller, and Cathcart, 1953).

No break in deposition occurs at the Triassic-Jurassic contact wherever it is observed within the area. Away from their common type locality in the Gabbs Valley Range, the formational boundary between the Gabbs Formation (Upper Triassic) and Sunrise Formation (Lower Jurassic), which is coincident with the systemic boundary, may not be recognizable. (*Silberling*)

24. *Column 19. Carson City-Reno area, Nevada.*—Isolated exposures of predominantly slaty rocks partly of Triassic age are present in the Pine Nut Range, Carson Range, Virginia Mountains (Gianella, 1936), and Nightingale Mountains (Smith and Guild, 1942) of western Nevada near the California border. In the Pine Nut Range the diagnostic Upper Triassic pelecypod, *Monotis subcircularis*, has been identified by S. W. Muller from black slates that grade upward into arenaceous and calcareous shales with "*Arietites*"-like ammonites of Early Jurassic age (Gianella, 1936, p. 37).

Undoubtedly some of the metamorphosed sedimentary rocks that occupy isolated exposures north of this area in Washoe County and in the western parts of Pershing and Humboldt counties are Triassic, but stratigraphic information is lacking. (*Silberling*)

25. *Column 20. West Humboldt Range, Pershing County, Nevada.*—The Star Peak and Koipato Groups were originally defined by the geologists of the Geological Exploration of the 40th Parallel (King, 1878; Hague and Emmous, 1877) from exposures in the West Humboldt Range. The same names were also applied by these authors to exposures in the East (Pah-Ute) Range, Tobin (Havallah) Range, Augusta Mountains, and New Pass (Desatoya) Range. Usage of the name Star Peak was extended by subsequent authors to other areas in northwestern Nevada until this name became more or less synonomous with the marine Triassic. More recently, largely as a result of studies by S. W. Muller and H. G. Ferguson, many new formational names have been introduced to express the stratigraphic relations and structural distribution of the Triassic rocks in northwestern Nevada. In relation to these more refined formational designations, the Star Peak is of doubtful utility even at its type locality.

The Star Peak Group was originally described by King (1878) and Hague and Emmons (1877) as alternating thick units of limestone and quartzite with a total thickness of about 10,000 feet. Triassic fossils were reported from near the top and base. Hyatt and Smith (1905, p. 21) recognized that the section described by the geologists of the 40th Parallel Exploration was exaggerated by faulting, but no revision of the Star Peak Group was attempted by these authors or by Smith (1914) in treating the paleontology of these rocks. The units characterized as "quartzites", originally included in the Star Peak, were first recognized as being volcanic by Ransome (1909, p. 33). Cameron (1939, p. 581) has shown that the "quartzite" (volcanic) units in the type section actually represent the top of the underlying Koipato group repeated by faulting. The type section of the Star Peak Group in the northeast part of the West Humboldt Range, as revised by Cameron, includes dominantly calcareous sediments of Middle Triassic age with an aggregate thickness of about 3500 feet. Current studies by the writer indicate that this section is equivalent to the Prida and part of the Natchez Pass Formations.

In Muttleberry Canyon, Upper Triassic rocks with *Monotis* are overlain by Lower Jurassic sedimentary rocks. Questionable Jurassic fossils are also reported by King (1878, p. 269) from the northwest tip of the Humboldt Range, and *Monotis* has been reported from hornfelses in the Eugene Mountains immediately north of the West Humboldt Range (Kerr, 1934, p. 14). The relationship of these Upper Triassic and younger rocks to the type section of the Star Peak is not yet known.

King (1878, p. 269) originally concluded that

"the Trias and Jura are perfectly conformable" in the West Humboldt Range, but later authors separated the Triassic from the overlying Jurassic sediments by an unconformity. Muller and Ferguson (1939, p. 1615) show that King's original conclusion was correct and that in the Muttleberry Canyon section "the *'Pseudomonatis' subcircularis* beds (Upper Triassic) can be followed upward into the fossiliferous beds of early Lower Jurassic age without any noticeable change in lithology or attitude."

Underlying the Star Peak Group is a thick section of dominantly volcanic rocks that King designated the Koipato Group. Like the volcanic rocks included by King in the Star Peak Group, these rocks were originally described as clastic sediments. King considered the age of the Koipato to be Middle and Early Triassic and was followed in this by subsequent authors until Wheeler (1939a, p. 107) reported the discovery of a diagnostic Permian fossil, *Helicoprion*, several thousand feet below the top of this unit. As the limestones near the base of the Star Peak Group (Prida Formation) contain fossils of medial Middle Triassic age, a disconformity is suggested between this unit and the underlying Koipato, although in the absence of further evidence the upper portion of the Koipato group may extend into the Lower Triassic. (*Silberling*)

26. *Column 21. Sonoma Range quadrangle, Nevada.*—The Triassic rocks of the Sonoma Range quadrangle, an area of about 3700 square miles in Pershing, Humboldt, and Lander counties, are separated by a major thrust fault into two distinct sequences which represent different lithologic and faunal facies of contemporaneous deposits (Muller, 1949). The lower-plate sequence or facies is restricted to the western half of the quadrangle, and the prevalence of clastic sediments in it is interpreted to be the result of shallow-water deposition. The Triassic rocks of the upper plate, on the other hand, contain a larger proportion of carbonate rocks and are presumably the product of off-shore deposition (Muller, Ferguson, and Roberts, 1951).

Both sequences overlie the Koipato Formation of Permian age with angular unconformity. The upper-plate facies, however, includes several thousand feet of sediments of Early Triassic age, whereas the oldest Triassic rocks present in lower-plate are of Middle Triassic age. Pre-Tertiary sedimentary rocks younger than Triassic have not been recognized within the limits of the Sonoma Range quadrangle.

A certain amount of lateral variation in lithology occurs within rocks of each thrust plate, and in the upper-plate sequence some of this variation is expressed by the formational designations applied. The limestones and shales of the Favret Formation, which are best developed in the southern exposures of the upper-plate facies, are supplanted to the northeast by coarse clastic sediments assigned to the China Mountain and Panther Canyon formations. Accompanying this lateral change in lithology, the China Mountain formation becomes the basal unit of the Triassic sequence in the northern exposures of the upper-plate facies (Muller, Ferguson, and Roberts, 1951).

The relationship of the Triassic rocks exposed in the Sonoma Range quadrangle to those of adjacent areas has not been adequately explored. To the southwest along the western flank of the New Pass Range the Lower, Middle, and Upper Triassic sediments are related to the upper-plate sequence in the Sonoma Range quadrangle. Further west, however, in the Clan Alpine and Stillwater ranges, the Triassic rocks exposed are lithologically and structurally more like the lower-plate sequence.

Upper Triassic rocks tentatively correlated with the Grass Valley, Winnemucca, and Raspberry formations of the lower-plate sequence have recently been described by Willden (1956) for northeastern and central Humboldt County, Nevada. In addition to these formations Willden also recognizes a new formational unit that may represent the interval between the Winnemucca and Raspberry formations and may be absent in the Sonoma Range quadrangle owing to faulting. (*Silberling*)

27. *Column 22. Southeastern Elko County, Nevada.*—Strata of Early Triassic age occur at many isolated localities throughout eastern Elko County (T. M. Cheney, personal communication), but most of these exposures have not been studied or described. The first record of Triassic rocks in this part of Nevada is provided by Smith (1932, p. 9–10), who reported fossils of the "*Meekoceras* zone" from a small exposure of limestone ". . . 3 miles west of the Phelan Ranch . . . on the east side of the Ruby Range." A similar occurrence is also described by Nolan (1935, p. 42) in the Gold Hill district of western Utah adjacent to Elko County.

Olive-drab to greenish-gray shales and subordinate limestone interbeds of Early Triassic age with a minimum exposed thickness of 3300 feet are described by Snelson (1955) from the southern Pequop Mountains. *Meekoceras* is re-

ported from the basal part of this section, which rests with unconformity upon rocks of Permian age. According to Scott (1954), strata referable to the Shinarump and Chinle formations overlie a similar section a few miles to the south. The sequence of Triassic units in the area between Currie and Spruce Mountain is described by Wheeler, Scott, and Thompson (1949) as being similar to the Woodside, Thaynes, Moenkopi, Shinarump, and Chinle formations. This section is said to grade upward into strata of questionable Early Jurassic age resembling the Nugget formation. In a subsequent paper (Wheeler, 1952) the formational designations applied to these strata were changed somewhat so that the Dinwoody replaces the Woodside Formation and the Timothy supplants the Moenkopi Formation. (*Silberling*)

28. *Column 23. Siskiyou Mountains, Oregon.*—Metamorphosed volcanic and sedimentary rocks of probable Triassic age in southwestern Oregon were described by Wells and Hotz (1941) and subsequently named the Applegate Group (Wells, Hotz, and Cater, 1949). The age of these rocks, based on fossil determinations by E. M. Kindle, was considered by Diller and Kay (1909, pp. 50–51) to be Devonian and Carboniferous. Upon restudy of Diller's collections, J. B. Reeside, Jr., determined the age of these rocks as definitely Mesozoic and probably Triassic. If the amphibolite gneiss exposed in the Kerby quadrangle is a metamorphosed facies of the Applegate Group, then the Galice Formation of Late Jurassic age rests with angular unconformity on the Applegate (Wells, Hotz, and Cater, 1949, p. 4). (*Silberling*)

29. *Column 24, Central Oregon, Crook and Grant counties.*—Triassic rocks were first reported in central Oregon by Packard (1928) from the headwaters of Beaver Creek and the South Fork of the John Day River in Crook and Grant counties. These rocks, from which a characteristic Triassic brachiopod fauna was obtained, were described as ranging from shale to conglomerate. Angular unconformities separate this section from the underlying sedimentary rocks of Paleozoic age and the overlying Jurassic rocks. This description was elaborated upon by Schenk (1934), who reported the Norian brachiopod *Halorella* from the upper part of a Triassic section 2700 feet in thickness and ammonites correlative with those of the *Tropites subbullatus* zone fauna of Northern California from another section at a somewhat lower horizon. At each locality the fossiliferous Upper Triassic strata are overlain by a thick sequence of unfossiliferous sediments which are in turn unconformably overlain by rocks of known Jurassic age. These sediments of uncertain age have been considered as belonging to the Upper Triassic but may include some strata of Early Jurassic age (Lupher, 1941, p. 228).

The Coyote Butte Formation of Permian age is the youngest Paleozoic formation recognized in the area and is unconformably overlain and overlapped by conglomerates and sandstones considered to be Triassic in age and equivalent in part to the sediments described by Schenk (Merriam and Berthiaume, 1943, pp. 158–159). (*Silberling*)

30. *Column 25. Wallowa Mountains, Oregon.*—The name Martin Bridge Formation was introduced by Chaney (1932, p. 4) for sedimentary and volcanic rocks of Late Triassic age, but no description of the locality or of the lithology of these rocks was provided. Previously Upper Triassic fossils had been reported from limestones in the Wallowa Mountains by Lindgren (1901, p. 581) and Smith (1927, p. 9–10). Ross (1938, p. 32–36) described the Martin Bridge as being dominantly calcareous and containing fossils of late Carnian age. The type locality was designated along Eagle Creek crossing the boundary between Wallowa and Baker counties. Volcanic deposits in the region, described as "Triassic (?) volcanic rocks," were considered to be equivalent to the Martin Bridge on the basis of their stratigraphic relations. Overlying the Martin Bridge Formation is a thick section of predominantly shaly rocks, designated the "younger Mesozoic rocks" by Ross, which were considered to be Jurassic and possibly Cretaceous in age. Smith and Allen (1941, p. 13) proposed the name Hurwal Formation for similar sediments in the northern Wallowa Mountains and showed that they gradationally overlie the Martin Bridge Formation and are probably wholly of Late Triassic age. The "Triassic (?) volcanic rocks" of Ross are included in the Hurwal formation. Definite evidence for Mesozoic sedimentary or volcanic rocks younger in age than late Triassic has not been discovered in the Wallowa Mountains.

Underlying the Martin Bridge Formation with "strong unconformity" is a group of altered sediments designated the "lower sedimentary series" by Smith and Allen (1941, p. 8). In the southern Wallowa Mountains sedimentary rocks with similar stratigraphic relationships, said to overlie the Clover Creek Greenstone of Permian age with no apparent

discordance and unconformably to underlie known Upper Triassic sediments, were assigned by Ross to the late Paleozoic. Smith and Allen (1941, p. 9, 56) consider the "Lower sedimentary series" to be at least in part of Middle Carnian age on the basis of fossils collected near the middle of the unit. This requires that a considerable amount of sedimentation, broken by a period of folding, took place during the middle and late Carnian, and consequently the structural and stratigraphic interpretation seems questionable.

A period of deformation and erosion preceding the deposition of the Martin Bridge Formation presumably is responsible for the absence of the "lower sedimentary series" in some localities where the Martin Bridge rests directly on the Clover Creek Greenstone (Smith and Allen, 1941, p. 10). (*Silberling*)

31. *Column 27, Northwestern Cascade Range, Washington.*—The dark argillites with subordinate sandstone, grit, and conglomerate of the Cultus Formation (Daly, 1912, p. 516–517) were assigned to the Triassic on the basis of fossils identified by T. W. Stanton as *Arniotites vancouverensis* Whiteaves? and *Aulacoceras*? sp. (resembling *A. carlottense* Whiteaves) and the lithologic resemblance to known Triassic rocks of Vancouver Island. As pointed out by McLearn (1953, p. 1216–1217) and Frebold (1953, p. 1232), the presence of *Arniotites*, if correctly identified, would place at least a portion of the Cultus Formation in the Lower Jurassic.

Crickmay (1930, p. 490) considers the Cultus Formation of Daly as representing alternating fault slivers of Carboniferous and Lower Cretaceous rocks and uses the names "Triassic volcanics" and "Slollicum Series" for rocks of supposed Triassic age in the Cultus Valley region. (*Silberling*)

32. *Columns 28, 29, 37–39. Arizona and Utah.*—The Moenkopi Formation of northern Arizona and southern Utah is assigned to Early and Middle Triassic time in the western and middle parts of this area and to Early Triassic in the eastern part. Its age is determined on the basis of marine invertebrates, amphibian and reptile bones, and vertebrate tracks.

The ammonite genus *Meekoceras* occurs near the base, in the Timpoweap or lowest of the six members that are present in northwestern Arizona and southwestern Utah; the genus *Tirolites* is in the Virgin or third member above the base in this area. Because *Tirolites* represents a late stage in Early Triassic time, yet is known only from the lower part of the Moenkopi formation in southwestern Utah, probably an unknown but considerable amount of the Moenkopi Formation in this area was deposited during Middle Triassic time (McKee, 1954, p. 11).

In northeastern Arizona and southeastern Utah the Moenkopi Formation is much thinner than it is farther west, apparently because transgression was from west to east, followed by regressions in the opposite direction. Thus, physical evidence indicates a span of time represented by the formation that was progressively less from west to east. Faunal evidence consisting of vertebrates (Welles, 1947, p. 286) and vertebrate trackways (Peabody, 1948, p. 413) indicates an Early Triassic age, possibly extending into Middle Triassic time, for the rocks of this area.

Unconformable above the Moenkopi Formation the Shinarump member of the Chinle Formation represents a continuous depositional sequence of continental deposits containing a well-developed vertebrate fauna. H. E. Gregory (1917) divided the Chinle above the Shinarump Member into four units designated from top to bottom by the letters A to D. Because of the general similarity between the fauna and that of the European Keuper (Lucas, 1901; 1904) and the presence of Late Triassic species of ostracodes (Cross, 1908, p. 107, 108), these rocks have long been placed in the Late Triassic. That none of the Chinle in this region is older than Late Triassic is shown, according to Camp (1930, p. 4), by the presence of *Stagonolepis* and associated forms in its lowermost beds in Arizona. Furthermore, he suggests that differences in the faunas of the upper part (divisions A and B of Gregory) and of the lower part (divisions C and D of Gregory) of the Chinle indicate that the strata containing them span a considerable part of Late Triassic time and that the upper part of the formation is younger than the Dockum group of Texas (Camp *et al.*, 1947, p. 8). Other students of the vertebrates, however, consider the Chinle and the Dockum Group essentially contemporaneous. (*See* Annotation 34.)

The red mudstone unit formerly referred to as division A of the Chinle (Gregory, 1917, p. 42), but now termed the Rock Point Member of the Wingate Sandstone (Harshbarger *et al.*, 1957), and the cliff-forming or Lukachukai Member of the Wingate are currently placed in the Upper Triassic. Furthermore, a red mudstone unit in the central and western parts of southern Utah and in the central part of

northern Arizona, called the Moenave Formation, rests on the Wingate Sandstone along its western margin and is considered of Triassic (?) age. Inasmuch as the Moenave and, where it is absent, the Wingate grade upward into the Kayenta Formation, which is listed as of Jurassic (?) age, the position of the boundary between the Jurassic and Triassic Systems in this area definitely is not known and must be placed at arbitrarily determined formational boundaries.

Evidence for the current age assignment of the Wingate Sandstone to Late Triassic time and of the Moenave to Triassic (?) time is based primarily on reptile and fish remains (Harshbarger *et al.*, 1957). The Triassic phytosaur *Machaeroprosopus* has been reported from the lower or Rock Point Member of the Wingate Sandstone in northeastern Arizona. The primitive crocodile *Protosuchus*, a new genus considered probably of Triassic age, has been found in the lower or Dinosaur Canyon Member of the Moenave Formation (Colbert and Mook, 1951, p. 151). Fish of the genus *Semionotus*, which has never been found above the Triassic, are known from the Moenave Formation in several localities in northern Arizona and southern Utah (Hesse, 1935; Schaeffer and Dunkle, 1950). (*McKee*)

33. *Column 30, Northwestern New Mexico.*—Throughout most of northwestern New Mexico only the Wingate Sandstone and the Chinle Formation are represented in the Triassic sequence (Harshbarger *et al.*, 1957). A thin edge of the Moenkopi Formation extends eastward from the Arizona State line as far as Fort Wingate but is absent north of that locality in New Mexico (McKee, 1954, p. 23). The Shinarump Conglomerate also extends as far east as Fort Wingate, where it is very thin according to Harshbarger (McKee, 1951, p. 88), but farther east it is not recognized.

The Chinle Formation in northwestern New Mexico is similar in lithology and thickness to the Chinle of the type area in Arizona. The Wingate Sandstone includes two units. The lower or Rock Point Member, formerly designated as member A of the Chinle Formation, extends eastward from Arizona far into the San Juan Basin area in the north but is more limited in the Gallup region in the south. Its Late Triassic age is indicated by the presence of the phytosaur *Machaeroprosopus*. The upper or Lukachukai Member is less widespread than the lower member in the northwestern corner of New Mexico but at the latitude of Fort Wingate extends considerably farther eastward (Harshbarger *et al.*, 1957). No diagnostic fossils have been found in it, but it has been assigned to a Triassic age because the overlying Moenave Formation farther west in Arizona contains fish of the genus *Semionotus*, considered to be Triassic. (*McKee*)

34. *Column 31, Central-Northern New Mexico.*—In the middle part of northern New Mexico the sequence of Triassic strata differs from that either east or west of it, owing to the addition of local conglomeratic units in the lower part of the Chinle Formation. Correlation of either the Agua Zarca Sandstone Member or the Poleo Sandstone Lentil of this area with lithologically similar units, such as the Shinarump Conglomerate and Santa Rosa Sandstone in adjacent areas, has not been established (Wood and Northrop, 1946), but all these conglomeratic sandstones are parts of a sequence formed during Late Triassic time. In north-central New Mexico (Kelly and Wood, 1946), the Correo Sandstone Member is recognized at the top of the sequence.

The vertebrate fauna of the Chinle Formation in the middle part of northern New Mexico has been discussed by Camp and others (1947, p. 8), who indicate a stratigraphic position similar to that of the type Chinle in Arizona, but extending somewhat higher into the Late Triassic than do strata of the Dockum Group as found in Texas. The relative abundance of the reptiles *Machaeroprosopus* and *Typothorax* in these two areas appears to have constituted the principal evidence of age (Camp, 1930, p. 3). Recent detailed studies by John Imbrie and E. H. Colbert of amphibians from the areas concerned, however, and increased knowledge of the distribution of other faunal groups inclines Colbert (personal communication, 1955) to the belief that "the Dockum is almost the exact equivalent of the Chinle." (*McKee*)

35. *Columns 32, 33. Northeastern and southeastern New Mexico.*—Triassic strata of eastern New Mexico, like those of central western Texas, are included in the Dockum Group and are commonly subdivided into formations, the uppermost of which is called the "Chinle." The extension of this name from the west and its restriction to the uppermost unit of the Triassic in this area is unfortunate, for here the name connotes a meaning different from that in the type locality. Its use is continued in this chart, in quotation marks, nevertheless, in deference to usage.

In northeastern New Mexico two formations

only, the "Chinle" and the underlying Santa Rosa Sandstone, are recognized, whereas the Tecovas, which forms the base farther east, is absent. In this area the upper part of the "Chinle" locally has been designated the Redonda Member (Dobrovolny and Summerson, 1947). Farther south in New Mexico and in adjoining parts of Texas, the Santa Rosa Sandstone is underlain by fine-grained red beds designated by Lang (1935, p. 264) the Pierce Canyon Formation and assigned by him (1937, p. 316) to the Triassic. It has been considered by Bates (1942, p. 46) to be an extension of the Tecovas, but recent petrographic studies by Miller (1955, p. 121) strongly suggest that it should be assigned to the underlying late Permian sequence. (*McKee*)

36. *Columns 34, 35. Central-Western and Panhandle Texas.*—The Dockum Group, Upper Triassic continental deposits bordering the Staked Plains of Texas and New Mexico, was described by Cummins (1890, p. 189) as the Dockum beds. Although generally accorded group status, its subdivisions, except possibly the Santa Rosa Sandstone, are all local. Terminology would be simplified if the Dockum were called a formation and its subdivisions were termed members, but in deference to usage it is here called a group, and its major parts formations.

Fossil vertebrates from the Dockum Group in Texas include forms similar to those of the Popo Agie Member of the Chugwater Formation and others similar to those in the lower levels of the Petrified Forest Member of the Chinle. All the Texas faunas appear less progressive than that obtained from the upper part of the Dockum Group in eastern New Mexico in beds below the Redonda Member, and from the upper levels of the Chinle in Arizona and north-central New Mexico (the *Machaeroprosopus tenuis* fauna). The Redonda Member in eastern New Mexico has yielded still more advanced phytosaurs, and dinosaur tracks. Its limited vertebrate fauna suggests an age later than that of the Petrified Forest Member of the Chinle Formation and is consistent with an age similar to that of the Wingate Sandstone.

The distribution of these successive faunas within the Dockum does not coincide with current correlations of formational subdivisions.

Primitive phytosaurs (*Paleorhinus* and *Angistorhinus*), found elsewhere only in the Popo Agie, occur in the Camp Springs Conglomerate near the base of the Dockum in Scurry County, Texas, in claystones several hundred feet above the base in Howard and Borden counties, and a few specimens have been found in Crosby County. These localities are above the so-called Santa Rosa Sandstone in this area but below the Medial sandstones of Drake. The *Paleorhinus* fauna is not known in association with more progressive phytosaurs of the type found in the Chinle. Localities containing this fauna are regarded as older than the Petrified Forest Member of the Chinle.

Phytosaurs, pseudosuchians, and amphibians from localities in Crosby County, from Palo Duro Canyon, and from the Canadian River Valley north of Amarillo, in Texas, all are closely similar to the fauna of the lower part of the Petrified Forest Member of the Chinle Formation near St. Johns, Holbrook, and Cameron, Arizona. The Canadian River localities are in shale typical of the Tecovas Formation, and there is good reason to regard the other localities mentioned as probably belonging to that unit. All are below the Medial sandstones of Drake and the Trujillo Sandstone of Gould. They are younger than the *Paleorhinus* beds to the south.

More advanced phytosaurs, comparable to those of the upper levels of the Chinle in the Petrified Forest area and to those of the Chama region in northern New Mexico, are present in the upper part of the "Chinle" portion of the Dockum near Tucumcari, New Mexico, but below the Redonda. No vertebrates so advanced as these have been found in Texas. They are associated with sandstones which may be correlatives of the Trujillo in Texas.

Vertebrates other than phytosaurs seem less useful for zonation of these deposits, although most of the fauna can be shown to indicate a Late Triassic age. Colbert and Imbrie (1956) found no significant differences between the metoposaurid amphibians from various localities in the Dockum; they considered the Dockum and Chinle specimens as belonging to a single population. *Eupelor* becomes rare in the upper levels of the Chinle Formation in Arizona and north-central New Mexico and is unknown from the Tucumcari region; at lower levels in the Dockum Group it is locally abundant. (*Gregory*)

37. *Column 34. Central-western Texas.*—The currently used threefold division of the Dockum into Tecovas Formation, Santa Rosa Sandstone, and "Chinle" Formation was introduced

by Adams (1929), who correlated the Santa Rosa Sandstone of eastern New Mexico with the Trujillo Sandstone of Gould (1907, p. 26–29), in the Canadian River valley and Palo Duro Canyon, Texas. Current opinion holds the Trujillo, Santa Rosa, and the "Middle sandstones" of Drake (1892) to be correlative.

The Tecovas Formation, described by Gould (1907, p. 23–26) from the Canadian River valley and best exposed in Palo Duro Canyon, was regarded by Adams (1929) as confined to the northeastern part of the Staked Plains. It is certainly absent farther west, in New Mexico, where the Santa Rosa Sandstone rests on San Andres limestone (Permian). From well cuttings in the Midland Basin Page and Adams (1940, p. 64) identified shales below the Santa Rosa Sandstone as Tecovas, and Bates (1942, p. 45) so identified beds exposed in the Pecos Valley that have also been called Pierce Canyon (Lang, 1935). Recent work, cited in Annotation 35, suggests that these beds may be of Permian age.

Along the eastern escarpment of the Staked Plains in Crosby, Garza, Borden, Howard, and Mitchell counties, Texas, several hundred feet of claystones and siltstones lies above the basal sandstones ("Santa Rosa Sandstone," Camp Springs Conglomerate) of the Dockum and below the Medial sandstones of Drake's sections (1892). In the southeastern counties these beds contain a vertebrate fauna more primitive than that of the type section Tecovas in the Canadian River valley. (*See* Annotation 36.) Sidwell (1945, p. 52) found that the mineralogical features of these sediments are more like those of the red shale member of the "Chinle" Formation near Montoya, New Mexico, than like those of the type Tecovas. No local term has been applied to these beds, and evidence for extending various named formations into this area is conflicting. Detailed field study will be necessary to determine the true stratigraphic relationships.

The Camp Springs Conglomerate is found locally at the base of the Dockum Group along the southeastern border of the Staked Plains. It has yielded fossil vertebrates similar to those of the Popo Agie (Langston, 1949). (*Gregory*)

38. *Column 35. Texas Panhandle.*—In its type area the Tecovas Formation (Gould, 1907, p. 23–26) rests directly upon Permian beds. Fossil vertebrates here are similar to those in the lower levels of the Petrified Forest member of the Chinle Formation. The overlying Trujillo sandstones (Gould, 1907, p. 26–29) have been correlated with the Santa Rosa sandstone of New Mexico but might also be equivalent to other sandstone channels at higher levels in the Dockum. Evidence of vertebrate fossils suggests that the Trujillo Formation is appreciably younger than the sandstones at the base of the Dockum which have been called Santa Rosa farther south, as in the Midland Basin of Texas. (*Gregory*)

39. *Column 36. Marathon Basin, Texas.*—In the Glass Mountain area of western Texas rocks composed largely of dolomite conglomerate but including interbeds of sandstone and limestone and lenticular layers of red shale have been named the Bissett Conglomerate by P. B. King (1927). This formation ranges up to about 800 feet in thickness, rests with discordance on various Permian and Pennsylvanian formations, and unconformably underlies Lower Cretaceous stata.

The age of the Bissett Conglomerate is very uncertain. It contains a poorly preserved flora originally assigned to the Upper Permian by David White but later considered by White as more probably Mesozoic (Sellards, 1932, p. 155). On this basis Lang (1935, p. 270) suggested that it be assigned to the "Lower Triassic," having "possible Moenkopi time equivalence."

Later examination of the Bissett flora by C. B. Read (*in* P. B. King, 1937, p. 110) led him to consider this formation older than the Dockum Group, though definitely of Triassic age. He states that the plant assemblage resembles a Middle Triassic flora of England. He lists the genera *Pelourdia, Zamites, Brachyphyllum, Voltzia,* and several fern fragments of obvious Mesozoic types.

A vertebrate bone, examined by E. C. Case (King, 1937, p. 110), on the other hand, resembles *Desmatosuchus* and suggests correlation of the Bissettt with the Late Triassic Shinarump and Dolores. Physical evidence supports this correlation. The limited distribution of the Bissett in an area bordering and immediately south of that containing the Dockum group of early Late Triassic age suggests that these rock units represent facies of the same depositional basin. For this reason the Bissett is tentatively shown as correlative with the Dockum group. (*McKee*)

40. *Columns 40–42, 48–54. Utah, Idaho, Montana, Wyoming.*—The Dinwoody Formation was named and defined by Blackwelder (1918, p. 425–426) from outcrops in Dinwoody

Canyon on the northeastern slope of the Wind River Mountains, near Dubois, Wyoming. The limits of the formation were defined by the Phosphoria Formation below and the bright red shales and siltstones of the Chugwater Formation above. Newell and Kummel (1942, p. 941) redefined the Dinwoody at the type locality to include only the dominantly silty strata between the Phosphoria and the top of the resistant siltstone about half way to the top of the original Dinwoody. The color boundary between the Dinwoody, as originally defined, and the red Chugwater was not a useful or natural boundary.

The Dinwoody in southeastern Idaho is 700 to 2400 feet thick and includes beds both older and younger than those found at the type locality in the Wind River Mountains, Wyoming.

The Woodside Formation was named by Boutwell (1907) for approximately 1000 feet of maroon and red shaly siltstone exposed in Woodside Gulch in the Park City mining district, northeastern Utah. The Woodside Formation overlies the Phosphoria Formation and underlies the Thaynes Formation. The red Woodside Formation and the nonred Dinwoody Formation form an intertonguing complex along an arcuate belt along southwestern Montana, the Idaho-Wyoming boundary, and westward in northern Utah, between the Phosphoria Formation below and the Thaynes Formation above. West of this arcuate belt the percentage of red Woodside rocks decreases rapidly, and east and south of it the red Woodside thickens rapidly; only a small part of the nonred Dinwoody Formation is present in central Wyoming, and none is found in the Wasatch Mountains and Uinta Mountains, Utah. The contact with the Phosphoria formation in southwestern Montana and southeastern Idaho is conformable. No physical evidence of a hiatus has been recognized. In southwestern and central Wyoming and the southern Wasatch Mountains, Utah, there is evidence of an unconformity between the Dinwoody and the Phosphoria formations (Kummel, 1954).

The Dinwoody Formation contains ammonite zones of the Otoceratan, Gyronitan, and Flemingitan ages. The lowest faunas of the Dinwoody contain *Ophiceras* cf. *O. greenlandicum*, *O.* (*Lytophiceras*) cf. *O. commune*, *O.* (*Glyptophiceras*) *nielseni*, *O.* (*Discophiceras*) *subkyokticum*, and *O.* (*Metophiceras*) *subdemissum*. These species are identical with or very similar to species in the *Ophiceras* beds (of Otoceratan age) of East Greenland (Kummel, 1954). In southwestern Montana a fauna of Gyronitan age is present, including *Prionolobus* n. sp. cf. *P. atavas* and *Koninckites* cf. *K. truncatus*. About 70 feet above this fauna are found ammonites including *Kymatites* n. sp. cf. *K. radiosum*, *Koninckites* n. sp. cf. *K. timorense* and *Xenodiscoides* cf. *X. involutus*, which suggest a Flemingitan age (Kummel, 1954). (*Kummel*)

41. *Columns 40, 41, 48–54. Utah, Idaho, Montana, Wyoming.*—Kummel (1954, p. 171) summarized the regional stratigraphic relations of the Thaynes Formation and is quoted here:

"The Thaynes formation was described by Boutwell (1907) from outcrops in Thaynes Canyon, Park City mining district, Utah. In the type region, the Thaynes formation consists of 1,190 feet of limestone, calcareous sandstone, sandstone, shale, and, in the middle, a red shale member. At the type locality, the Thaynes is underlain by the Woodside formation and overlain by the Ankareh formation. The Thaynes formation is found over a wide area in northern Utah, eastern Idaho, western Wyoming, and southwestern Montana. Along its southern, eastern, and northern margins it inter-tongues with the red Ankareh formation or the Chugwater formation. The thickest development of the Thaynes is in the Fort Hall Indian Reservation. No red beds are present in this section. The Thaynes in southeastern Idaho is differentiated into several lithologic units, which can be traced over a wide area. In southwestern Montana, in most of western Wyoming, and in northern Utah, the lithology of the Thaynes formation is more homogeneous.

"The base of the Thaynes formation throughout eastern Idaho is a prominent limestone bed containing an abundant ammonite fauna characterized by *Meekoceras*. This same fauna is found in the Wasatch Mountains, Utah, and in southwestern Montana. In western Wyoming, the limestone with *Meekoceras* is present in the Salt River Range and the Sublette Ridge. In the Wyoming Range and Teton Range, the lower limestone with *Meekoceras* is not present and that horizon is occupied by red beds of the Woodside formation. The eastward thinning of the Thaynes formation in western Wyoming takes place by lateral change into the lithology of the underlying Woodside formation and the overlying Ankareh formation. The Thaynes formation completely disappears east of the Wyoming Range. In the Fort Douglas area of northern Utah, the Thaynes formation consists of a thick, normal marine succession. Eastward along the Uinta Mountains, the Thaynes formation thins and passes rapidly into the red Ankareh and the Woodside formations. Thus, the regional relationship of the Thaynes formation is very similar to that of the Dinwoody formation except that the boundary area between the red facies and the marine facies goes south from eastern Idaho through the Wasatch Mountains, Utah, rather than westward in northern Utah as for the Dinwoody formation.

"The Thaynes is the most fossiliferous formation of Triassic age in the Middle Rocky Mountains.

Several distinct ammonite zones are recognized, which are present only in the thicker geosynclinal sections. In the areas where the Thaynes thins and red beds intertongue, the ammonites are generally not present. Smith (1932) recognized three ammonite zones in the Thaynes formation; these are in ascending order the *Meekoceras*, *Tirolites*, and *Columbites* zones. Mathews (1929) described from the Fort Douglas area of Utah, a fauna which is present just above the *Meekoceras* zone that is referred to the *Anasibirites* zone. This zone is also present in southeastern Idaho. A new and highly important zone, the *Prohungarites* zone, is present above the *Columbites* zone in the upper part of the Thaynes formation in the Bear River Range."

In part of eastern Idaho the Thaynes Formation is interpreted to include at the top the Timothy Sandstone Member and beneath it the Portneuf Limestone member, into which the Lanes Tongue of the Ankareh Formation passes from the southeast. It has been proposed to correlate the Portneuf Member with the Upper Triassic Alcova Limestone member of the Chugwater Formation of Wyoming, but it is here considered that the Alcova is a younger limestone and that the beds to the top of the Timothy Member are part of a sequence of Lower Triassic deposits. (*Kummel*)

42. *Columns 40–42, 49–51, 54. Utah, Idaho, Wyoming.*—Kummel (1954, p. 179) summarized the regional stratigraphic relations of the Triassic rocks included between the Thaynes Formation and the Nugget Sandstone and is quoted here.

"A variable sequence of beds that are generally red lies between the top of the Thaynes formation and the base of the Nugget sandstone throughout western Wyoming, southeastern Idaho, and northern Utah. Boutwell (1907) called these beds, in the Park City district, [Utah,] the Ankareh formation. Although they are still so defined in western Wyoming, they have since been subdivided and named differently in other areas. There has been a great deal of confusion in nomenclature and correlation of these formations.

"In the original definition of the Ankareh, Boutwell (1907, p. 453) stated that only part of this formation occurs in the Park City district and that the uppermost part was marked by a prominent, massive, white sandstone member. Boutwell (1912) later restricted the Ankareh and included the upper 500 feet of white sandstone with some intercalated reddish shale in Veatch's Nugget formation as redefined by Gale and Richards (1910). A similar procedure was followed by Mathews (1931) and Boutwell (1933) for the central Wasatch Mountains. These authors thus considered the Ankareh to include the red strata from the top of the Thaynes to a prominent "basal conglomerate" (Gartra grit of Thomas and Kreuger, 1946), which is associated with red shales they thought to be part of the Nugget.

"Thomas and Kreuger (1946) reviewed in detail the literature on these post-Thaynes Triassic rocks for the central Wasatch and Uinta Mountains. They proposed a threefold division of this interval into Ankareh formation, Gartra grit, and Stanaker formation. Williams (1945) proposed using the Idaho terminology, Ankareh formation, Higham grit, and Wood shale for the formations in the central Wasatch Mountains and western Uinta Mountains, and the Colorado Plateau terminology, Red Wash formation, Shinarump conglomerate, and Chinle formation, for those in the eastern Uinta Mountains.

"Baker (1947) included in the Ankareh all the red strata between the Thaynes and the Nugget formations in the region around Provo, Utah. A. E. Granger (personal communication) followed the same procedure in mapping the area just east of Salt Lake City.

"For southeastern Idaho and adjacent parts of Utah and Wyoming, Gale and Richards (1910) used the name Ankareh for red shales lying between the Thaynes and the restricted Nugget formation of Veatch. Schultz (1914) used the name Ankareh for red shale and sandstone lying between the Thaynes and the Nugget formations in Lincoln County, Wyo. This usage is still followed in western Wyoming (W. W. Rubey, personal communication.).

"In southeastern Idaho the thickest sequence of post-Thaynes Triassic rocks is found in the Fort Hall Indian Reservation. Mansfield (1927) recognized the following formations: Timothy sandstone, Higham grit, Deadman limestone, and Wood shale. These units thin and tongue out eastward toward Wyoming.

"As now used the name Ankareh formation is applied to different parts of the post-Thaynes sequence in the Middle Rocky Mountains. In western Wyoming it is applied to all the strata between the Thaynes formation and the Nugget sandstone. In Utah it is applied only to the beds from the Thaynes to the Gartra grit. In Idaho the name Ankareh was abandoned by Mansfield. However, the Gartra grit is probably equivalent to the Higham grit, and the Higham grit is present in western Wyoming in the Ankareh formation. To simplify and clarify the nomenclature of these post-Thaynes Triassic rocks, the Ankareh formation as originally defined by Boutwell to include all the strata between the Thaynes and Nugget formations is here applied in northern Utah and western Wyoming, and all the various formations that have been proposed for parts of this sequence reduced to member rank. In the eastern Uinta Mountains, the Colorado Plateau nomenclature is used, following Kinney and Rominger (1947). In southeastern Idaho, the Timothy sandstone is now considered to be the uppermost member of the Thaynes formation. The Higham grit and Deadman limestone are ranked as independent formations, and the Wood shale is considered a westward extending tongue of the red Ankareh formation as used in Wyoming."

The Nugget Sandstone is assigned in this chart somewhat arbitrarily to the Triassic. The typical Nugget of southwestern Wyoming has not yielded fossils nor have other units correlated with it. The Nugget underlies Middle Jurassic marine beds, rests on nonmarine beds assigned to the Upper Triassic, and has been

placed heretofore in the Jurassic chiefly because of an assumed equivalence to the Navajo Sandstone of Utah and Arizona. The Navajo itself has provided little faunal evidence for its age but is assigned to the Jurassic because of local intertonguing with the marine Carmel Formation and because it is later than the Kayenta Formation, which in its turn is now assigned to the Jurassic with doubt because of a seeming conflict in the interpretation of the pertinent vertebrate fossils. (*See* discussion of Rhaetian deposits by Colbert and Gregory.) In central Wyoming beds assigned to the Nugget Sandstone interfinger with the upper part of the Popo Agie Member of the Chugwater Formation and with the Jelm Formation, and the lower part of the Popo Agie contains early Late Triassic vertebrates. (*See* Annotation 47.) This would not in itself prove that all the Nugget is of Triassic age, but it is considered here most useful to regard it as such. (*Kummel*)

43. *Column 43. Southwestern Colorado.*—In southwestern Colorado rocks of Triassic age have been named the Dolores Formation for typical exposures in the valley of the Dolores River (Cross, 1899). These strata are overlain by the Entrada Sandstone of Late Jurassic age and underlain by red beds of the Cutler Formation of Permian age. Near Ouray the Dolores Formation rests with angular unconformity on Permian strata (Cross and Howe, 1905).

The Dolores formation is said to include equivalents of the Kayenta Formation, Wingate Sandstone, and Chinle Formation (Baker *et al.*, 1936, p. 34, 39). It contains fossil plants and vertebrate remains, including parts of phytosaurs which indicate a Late Triassic age, but as yet no fossils have been reported that show its exact position within this series. Lithologically it resembles parts of the Glen Canyon Group, especially the Rock Point Member of the Wingate Sandstone, but differs from strata of the type Chinle Formation. Thus, although its lower part may correspond in time (but not in facies) with the type Chinle, the Dolores Formation is here tentatively indicated as equivalent to the lower or Triassic part of the Glen Canyon Group. (*McKee*)

44. *Column 44. Northeastern Colorado.*—Lee (1927, p. 14, 33, 37, 40) tentatively referred to the Jelm Formation an orange sandstone up to 125 feet thick lying between the Lykins Formation and the Upper Jurassic. It has been the practice since to call this unit the Jelm, though paleontologic evidence of its age is lacking.

Strata forming the upper part of the Lykins Formation along the northern part of the Front Range and eastward in the subsurface of northeastern Colorado have long been considered of Triassic age (Eldridge, 1896, p. 57). Evidence in the form of fossils for this age assignment is lacking. It is based partly on lithologic similarities between the Lykins and established Lower Triassic red beds to the north and west and partly on stratigraphic position.

The position of the boundary between the supposed Triassic red beds and those of Permian age in the Lykins Formation of northeastern Colorado is not known. Some geologists arbitrarily place it at the top of the "upper crinkly" limestone because that unit is readily recognized and mapped (LeRoy, 1946, p. 44), but this limestone (the Glennon of LeRoy), according to L. Broin (Personal communication), is correlative with and traceable into the Forelle Limestone of eastern Wyoming which is not considered to be as young as the top of the Permian. His recent work suggests that changes in lithology of the red beds, above the uppermost thin limestone in the Lykins and 50 to 100 feet above the "upper crinkly," mark a more likely boundary. This limestone is tentatively correlated with the uppermost tongue of the Phosphoria to the north and northwest in Wyoming.

In the central part of the Front Range in Colorado, because red beds above the "upper crinkly" limestone in the Lykins formation are thin or lacking, Triassic strata almost certainly are absent. (*McKee*)

45. *Columns 45, 46. Southeastern Colorado and northwestern Oklahoma.*—Triassic rocks of southeastern Colorado and western Oklahoma appear to be northerly and easterly extensions of strata assigned to the Dockum Group in northeastern New Mexico. Like the New Mexican strata, they apparently represent parts of the Upper Triassic, as indicated by vertebrate fossils reported by Stovall (1943, p. 50) from Cimarron County, Oklahoma, and by T. G. McLaughlin (1954, p. 85) from Baca County, Colorado. These fossils include the phytosaur *Machaeroprosopus* and the amphibian *Eupelor* (includes *Buettneria* of authors).

The name Dockum Group is commonly used to designate all rocks of Triassic age throughout much of the southern High Plains, but locally upper parts are known as the Sloan

Canyon Formation and the Sheep Pen Sandstone (Parker, 1933, p. 41). (*McKee*)

46. *Column 47. Southwestern Kansas.*—Red beds outcropping locally in Morton County, Kansas, and found more extensively in the subsurface have been assigned tentatively to the Dockum Group (Moore, 1951, p. 29) on the basis of lithology and stratigraphic position. (*McKee*)

47. *Columns 55–66. Wyoming.*—As a result of surface, subsurface, and electric-log studies of the Triassic rocks of Wyoming, in connection with assembling data for this correlation chart, the following points appear worthy of emphasis:

(1) The relationship of the thick marine Lower Triassic Thaynes Limestone sequence in western Wyoming to the Alcova limestone Member of the Chugwater Formation of central Wyoming has long been the subject of controversy. Recent wells drilled in Teton County, Wyoming, and Teton County, Idaho, suggest that the Thaynes Limestone intertongues with red beds below the Alcova, and that the Alcova either is a younger limestone or, less probably, intertongues with the topmost part of the Thaynes. In this chart the Alcova Limestone is considered to be of Late Triassic age.

(2) The red-bed sequence below the Alcova Limestone in most of Wyoming contains many thin but very widespread units that form distinctive patterns on electric logs. By means of these characteristic records, entire sequences of units can be correlated for distances of as much as several hundred miles.

(3) The rocks above the Alcova Limestone Member have fewer persistent zones than those below, and the rapid lateral facies changes from shale and siltstone to sandstone make correlation difficult. Conditions of deposition apparently were much less stable after Alcova time than before it.

(4) The Nugget Sandstone, originally established in southwestern Wyoming (Veatch, 1907), but modified by later students to include only the uppermost part of the original unit (Mansfield, 1927), has been considered to be Lower Jurassic by most recent writers. However, sandstones assigned to the Nugget in central Wyoming intertongue eastward and southward with the Popo Agie Member of the Chugwater Formation, which in its lower beds contains an early Late Triassic vertebrate fauna. Possibly some part of the Nugget Sandstone may be of Early Jurassic age, but, if so, in the compiler's judgment new criteria will have to be discovered to distinguish the Jurassic part from that which intertongues with the Popo Agie. In this compilation the Nugget is considered to be entirely of Triassic age.

(5) The relationship of the Jelm Formation in the Laramie Basin to the Nugget Sandstone of western and central Wyoming has been a point of controversy for many years. As a result of recent drilling and new electric-log data, as well as surface studies by several graduate students, directed toward solving this problem, the Jelm Formation, which is largely a sandstone facies in the type area, is known to intertongue northwestward with claystones and shales of the Popo Agie Member of the Chugwater Formation.

(6) The Dinwoody Formation in southwestern Wyoming contains greenish-gray dolomitic siltstone beds with distinctive electric-log curves that can be recognized in the red siltstone and shale sequence of the lower part of the Chugwater Formation in central Wyoming. (*Love*)

48. *Column 55. Rock Springs Uplift, Wyoming.*—The section below the Alcova Limestone Member of the Chugwater Formation can be correlated very precisely with the lower part of the Chugwater and the Dinwoody Formation in central Wyoming, by means of electric logs and sample studies of well cuttings. At least a dozen control horizons can be recognized, and they indicate that the same depositional layers that are green sandstone in the Dinwoody formation in the Rock Springs area are red siltstones in the lower part of the Chugwater Formation in central and southeastern Wyoming. The Popo Agie Member contains abundant analcite clusters and ocher claystone. The Nugget is essentially a solid sandstone mass with no shale partings except in the lower 75 feet. (*Love*)

49. *Column 56. Rawlins area.*—Rocks of Dinwoody age are represented by red beds in the lower part of the Chugwater Formation. This can be demonstrated by correlating resistivity curves on electric logs from this area northwestward to where the rocks grade to greenish-gray dolomitic siltstones of the Dinwoody Formation. The lower part of the Nugget Sandstone has graded laterally into a red-bed facies of the Popo Agie Member. This can be demonstrated by integrated studies involving both electric logs and surface sections. (*Love*)

50. *Column 57. Laramie Basin, Wyoming.*—

Although readily determinable on electric logs, the base of the Triassic is difficult to recognize on outcrops in parts of this area, for red beds, gypsum, and anhydrite occur both above and below the contact. The Permian rocks contain more dolomite, limestone, gypsum, and anhydrite. The Alcova Limestone Member of the Chugwater Formation disappears, apparently by nondeposition, in the northern part of the Laramie Basin, but throughout the remainder of the basin to the south the horizon of the Alcova is recognizable in electric logs. The Jelm Formation directly overlies this horizon and consists largely of sandstone in the type locality. Vertebrate fossil fragments occur in the Jelm Formation, but as yet none has been sufficiently well preserved to be useful in age determination. Northwest of the Laramie Basin the sandstones of the Jelm Formation intertongue with red shales and siltstones of the Popo Agie Member. In the uper part of the Jelm Formation is a light-colored sandstone, locally called "Nugget" or Troublesome Creek Member (Hubbell, 1956), that appears to intertongue eastward with the upper part of the Popo Agie Member of the Chugwater Formation. Between the Laramie Basin and the Rawlins Uplift, the thick section of post-Alcova Triassic red beds contains very little sandstone and is directly overlain by Upper Jurassic rocks. It is difficult to correlate intertonguing sandstones on both sides of this area of siltstone and shale red-bed facies. East of the Laramie Basin the Jelm Foramtion has been removed by pre-Upper Jurassic erosion. (*Love*)

51. *Column 58. Chugwater Creek, Wyoming.*—Darton's type Chugwater Formation contained only a small part of the sequence now called Chugwater in areas to the west and northwest. The Alcova Limestone and younger rocks were removed by pre-Upper Jurassic erosion. The basal part of the type Chugwater is a limestone, gypsum, and red-shale sequence that is now known to be Permian and has been split into the Satanka Shale, Forelle Limestone, Glendo Shale, and Minnekahta Limestone. As currently used in this area, the Chugwater Formation is the red-bed sequence overlying the Minnekahta Limestone. On electric logs, however, a Permian-Triassic contact consistent with that to the west is picked slightly higher in the red-bed sequence. (*Love*)

52. *Column 59. South margin of Yellowstone Park, Wyoming.*—The Dinwoody and Chugwater Formations show the same facies present in areas to the southeast and have not been satisfactorily correlated with the more marine sequence to the southwest in eastern Idaho. Therefore, the central Wyoming terminology is extended to include the Triassic rocks in southern Yellowstone Park. The upper part of the Popo Agie Member and all of the Nugget Sandstone have been removed by pre-Middle Jurassic erosion. (*Love*)

53. *Column 60. Jackson Hole, Wyoming.*—The Chugwater Formation is a red-bed facies typical of that in areas to the east and southeast and is not like the marine gray limestone sequence to the southwest. Despite this, some writers have extended the southwestern terminology into Jackson Hole, but for this chart the central Wyoming terminology seems more appropriate. Vestiges of two tongues of the Thaynes Limestone are present below the Alcova Limestone in the extreme southern part of the area. The Nugget sandstone thins northward very rapidly by pre-Middle Jurassic erosion and is absent in the central and northern parts of Jackson Hole. (*Love*)

54. *Column 61. Western part of Wind River Basin, Wyoming.*—This is one of the most extensive and clear-cut Triassic sections in Wyoming and includes type localities of four formations or members. The Crow Mountain Sandstone Member of the Chugwater Formation thins and disappears westward toward Jackson Hole, thickens eastward, and thins but is recognizable southward for about 75 miles. The Alcova Limestone Member is locally absent in the northwestern part of the Wind River Basin, and in those places the Crow Mountain extends for a few feet below the horizon of the Alcova. The Popo Agie Member contains a Late Triassic vertebrate fauna. (*Love*)

55. *Column 62. Casper-Alcova area, Wyoming.*—The Permian-Triassic contact is placed at the top of the Ervay Limestone Tongue of the Phosphoria Formation, even though there are red beds and anhydrites both above and below it. A nothosaur was collected from the Alcova Limestone Member of the Chugwater Formation and has been compared with Middle or Upper Triassic forms by some workers. The Crow Mountain Sandstone Member is well developed, but most younger rocks have been removed by pre-Upper Jurassic erosion. The basal sandstone of the Sundance Formation resembles both the Nugget and Crow Mountain Sandstones and may well have been derived from them. It is in contact with the

Crow Mountain Sandstone Member and has been called Nugget by many geologists, but appears to be continuous with the basal ammonite-bearing sandstone of the Sundance formation in the Glendo area to the east. (*Love*)

56. *Column 63. Glendo area, Wyoming.*—The Permian-Triassic contact is difficult to pick on surface outcrops but is moderately clear cut in electric logs and is commonly placed at the top of a thick gypsum and anhydrite sequence interbedded with red shale and limestone, overlying the Minnekahta Limestone. The Alcova Limestone Member and younger Triassic rocks have been removed by pre-Upper Jurassic erosion. (*Love*)

57. *Column 64. Southern Bighorn Basin, Wyoming.*—The Dinwoody Formation becomes gypsiferous and intertongues eastward with red beds. The Crow Mountain Sandstone Member of the Chugwater Formation thickens eastward but thins northward. The Nugget Sandstone is present locally in a few places along the extreme southern margin of the basin, but farther north both it and the underlying Popo Agie Member of the Chugwater Formation have been removed by pre-Middle Jurassic erosion. (*Love*)

58. *Column 65. Northern Bighorn Basin, Wyoming.*—Dinwoody Formation intertongues eastward with red beds but is readily distinguishable in electric logs. The Alcova Limestone Member of the Chugwater Formation and younger Triassic rocks were removed eastward by pre-Middle Jurassic erosion. (*Love*)

59. *Column 66. Western Powder River Basin, Wyoming.*—The Permian-Triassic contact is placed at the top of a limestone, gypsum, anhydrite, and red-shale sequence on surface outcrops. This is approximately the same contact that is picked on electric logs and that can be followed westward to the contact between the marine Phosphoria and Dinwoody Formations. The Alcova Limestone Member of the Chugwater thins and disappears, apparently by nondeposition, and where it is absent the Crow Mountain Sandstone Member extends for 10–20 feet below the horizon of the limestone. The sandstone is thick and widespread in this area. Younger Triassic rocks have been removed by pre-Middle Jurassic erosion. (*Love*)

60. *Column 67. Western margin of Black Hills, Wyoming.*—The Permian-Triassic contact is the subject of controversy. Some geologists put it at the top of the Minnekahta Limestone, others at the top of the dolomite, anhydrite, and red-shale sequence, and still others within the latter sequence. Because of the lack of fossils, the only possibility of settling the dispute is to make a very detailed correlation westward to marine Phosphoria and Dinwoody in the Bighorn Basin, by using electric logs, sample studies, and outcrop studies. The studies of this type that have been made so far suggest to the writer that the probable contact may be at the top of the dolomite and anhydrite part of the sequence.

The Alcova Limestone Member of the Chugwater Formation and younger Triassic rocks have been removed by pre-Middle Jurassic erosion. There are a few localities where red sandstone, similar to the Nugget, Crow Mountain, Jelm, and several other Triassic sandstone zones, overlies the Spearfish Formation with a sharp lithologic break and unconformably underlies the Gypsum Spring Formation in some places and the Sundance Formation in others. It seems to the writer more likely that this sandstone may be a lateral equivalent of the Crow Mountain Sandstone Member of the Chugwater than of the Nugget Sandstone. It rests on rocks stratigraphically well below the Alcova in sections to the west, so if the sandstone is either Nugget or Crow Mountain there would be a considerable unconformity within the Triassic sequence in the Black Hills area. (*Love*)

61. *Columns 68, 69. Eastern Black Hills, South Dakota, and Williston Basin, North Dakota.*—Rocks of Triassic age occur in the subsurface of the Williston Basin. The Black Hills of South Dakota exhibit the only outcrop of Triassic rocks around the margin of the basin. Therefore, the likely age of various units of the rocks exposed there (Spearfish Formation) has been tentatively established by subsurface correlation across the Powder River Basin to the Big Horn Basin where paleontological data have been secured. It is thought that a reasonable lithologic correlation from the Black Hills outcrop to the subsurface of the Williston Basin would provide an accurate interpretation, as these red beds in the subsurface have not yielded any fossils.

The Spearfish Formation has been defined by Darton (1899, p. 387) to be the red shale, sandstones, and gypsums above the Minnekahta Formation of Permian age and below the Sundance of Jurassic age. Present-day correlation indicates that rocks of Permian age are included in Darton's Spearfish. The following lithologic units, as defined in a surface section

at Buffalo Gap, South Dakota, can be correlated westward into the Powder River Basin and northward into the Williston Basin:

Unit 1: This unit conformably overlies the Minnekahta Limestone. It is composed of 85 feet of red shale, green shale, and gypsum with a minor amount of dolomite. The green shale, found in the top of the unit, is thought to be correlative with the green shale that is characteristic of the uppermost Phosphoria of the Big Horn Basin. Subsurface correlation eastward across the Powder River Basin of Wyoming has established the Phosphoria equivalent in the subsurface on the west flank of the Black Hills. This correlation indicates that the basal portion of the "Spearfish" is of Permian age.

Unit 2: Conformably overlying Unit 1 is 35 feet of gypsum, dolomite, and red shale. Unit 2 is considered representative of a Dinwoody equivalent which can be carried in the subsurface to the west flank of the Black Hills. Some geologists may desire to place the Permian-Triassic boundary at the top of Unit 2, but the compilers consider the subsurface correlation with the Dinwoody formation of the Big Horn Basin to be of sufficient strength to make the top of Unit 1 the most likely level of Permian-Triassic contact.

Unit 3: This unit consists of 118 feet of red shale and gypsum. Three prominent gypsum beds appear equivalent to a tripart subsurface correlation unit that can be carried widely by means of electric logs. Correlation can be carried across the Powder River Basin into the Chugwater formation of the Big Horn Basin.

Unit 4: The uppermost unit of the "Spearfish" of the Black Hills is 80 feet of orange-red siltstone grading to fine sandstone, which is a coarser type of clastic than the underlying sediments. This unit is recognizable in the Big Horn Basin where the Middle Jurassic Gypsum Spring Formation overlies it. At the base of the siltstone, there occurs in outcrop a patch conglomerate composed of brown chert, dolomite, and limestone pebbles. The Upper Jurassic Sundance Formation overlies unit 4 unconformably in the Buffalo Gap surface section. Farther north, in the vicinity of Spearfish, South Dakota, the Gypsum Spring Formation overlies this siltstone unit.

In the subsurface of the Williston Basin above the Minnekahta and below the Jurassic, the following lithologic members can be recognized in descending order: A—Upper salt member; B—Siltstone member; C—Middle salt member; D—Shale member.

These subsurface members will be discussed in relation to the Black Hills outcrop units. Owing to poor samples, the lithology in certain drilled wells is inexact, but by the selection of instances of good samples a reasonably good interpretation can be made.

The subsurface shale member D is composed of red and green shale. The lower green-shale portion of the unit is considered equivalent to surface Unit 1, thus of possible Phosphoria age. The upper red-shale portion of subsurface member D is considered equivalent to surface Unit 2. A few occurrences of limestone above the Minnekahta have been encountered in subsurface member D, which limestone may be equivalent to the Ervay or Forelle Limestone Tongues of the upper part of the Phosphoria that are present in southern Wyoming. Therefore, the lower portion of the shale member D appears to be Permian in age, and some part of the upper portion to be Triassic; member D contains somewhere within its upper portion the Permian-Triassic boundary.

The middle salt member C represents a very restricted environment, with conditions not commonly found in the Triassic of the Powder River and Big Horn basins. This member is equivalent to Unit 3 of the Black Hills and may contain a portion of surface Unit 2. However, in general it is thought that surface Unit 2 is part of member D of the subsurface. Much evidence of evaporitic environment is found in Unit 3 of the Black Hills section.

The subsurface siltstone member B is correlative with surface Unit 4. The local occurrence of a conglomerate at the base of member B has been established. In the central portion of the Williston Basin, siltstone member B conformably overlies and underlies the adjacent rock units. On the margins of the basin, siltstone member B lies with marked unconformity on rocks as old as Mississippian. There has been much dispute concerning the age of member B beyond the subsurface limits of members C and D. The increased thickness of siltstone member B in southern Saskatchewan and the apparent conformity with the overlying Jurassic in that area have been the principal reasons for the assignment by some geologists of member B to a Jurassic age along the International boundary.

The upper subsurface salt member A is restricted to west-central North Dakota in the area east, west, and south of the Nesson anticline. The age of this salt is uncertain, but favor is given to the Triassic. The position of

its development and the lithology suggest to the authors a closer affinity to the underlying Triassic than to the overlying Jurassic. No surface equivalent to salt member A is known.

Zieglar in a recent paper (1955) has used for the subsurface units of the Williston Basin the following names: Unit D, Spearfish Formation, restricted; Unit C, Pine Salt; Unit B, Saude Formation; Unit A, Dunham Salt. Francis (1956) used for the pre-Gypsum Spring strata of Mesozoic age of the Williston Basin the designation "Jura-Triassic red beds;" he drew the boundaries on lithologic criteria and discarded the designation Spearfish as implying more than is known.

No paleontological data have been forthcoming from the Williston Basin to establish the age of the rocks in question. The general environment, as typified by the lithology, indicates that a meager fauna, if any, existed. Sedimentation studies might help greatly in establishing unit correlation.

(*Lewis and Hadley*)

62. *Columns 70–74. Alabama, Georgia, Florida.*—Many of the data relating to the subsurface Triassic(?) rocks in the southeastern Coastal Plain are taken from U. S. Geological Survey Circular 91 (Applin, 1951) and from the writer's unpublished information gained from later laboratory studies; all relevant published data have been considered, and much of it utilized.

A sequence of terrestrial or marginal clastic rocks tentatively classified as Triassic were penetrated in 12 scattered oil-test wells in southeast Alabama, southwest Georgia, and north-central Florida. Lithologically, the subsurface sedimentary rocks are closely similar to rocks of the Triassic Newark Group that crop out at various localities along the Atlantic seaboard from Massachusetts to North Carolina; also, like the Newark Group, intrusions and flows of diabase and basalt cut the Triassic (?) strata in several wells (Southeastern Geological Society, 1949; Applin, 1951, p. 15–17, 26–27). The occurrence in the subsurface of Triassic(?) rocks in Georgia, Florida, and Alabama suggests the possible southwestward extension of the zone of Triassic sedimentation, faulting, and volcanism from North Carolina into the subsurface of the southeastern Coastal Plain. The available data from the scattered wells do not clearly define the structural relationships of the Triassic(?) strata in Alabama and Georgia. In north-central Florida, however, Triassic(?) strata wedge out against the southwest flank of the Peninsular Arch (Applin, 1951, p. 3, 5), the dominant subsurface structural feature of the peninsula as far south as the latitude of Lake Okeechobee.

Cores and cuttings from the different wells (Table 4; Fig. 6) show that the buried Triassic(?) sedimentary rocks are composed, in general, of hard, dark-red and greenish-gray and mottled, micaceous shales irregularly interbedded with fine-to-coarse-grained, poorly sorted, white, gray, and red sandstones, which, in places, are conglomeratic and highly arkosic. For the most part, the strata are unfossiliferous, but according to John B. Reeside, Jr., U. S. Geological Survey, a core from the depth of 6144–6149 feet in the Humble Oil & Refining Co.'s Hodges well 1 (sec. 12, T. 5 S., R. 6 E.), Taylor County, Florida, contained specimens of the genus *Cyzicus* (*"Estheria"*), a long-ranging form that is among the most common fossils of the Newark Group. Four wells in the Alabama-Georgia-Florida area terminated in Triassic(?) sedimentary rocks, and four terminated in diabase that is overlain by Triassic(?) strata (Applin, 1951, p. 15–17, 26–28); in four other wells the Triassic(?) sedimentary rocks unconformably overlie strata of early Paleozoic age. The upper boundary of the Triassic(?) sedimentary sequence is not sharply defined, and the available subsurface data indicate that the Triassic(?) strata are overlain by an unfossiliferous clastic facies of undifferentiated Lower Cretaceous rocks.

East of the area in southeast Alabama, southwest Georgia, and north-central Florida that is occupied by the Triassic(?) sedimentary rocks, nine widely scattered wells penetrated flows and intrusions of diabase that are tentatively classified as Triassic (Applin, 1951, p. 15–17, 26–27). Three wells in southeast Georgia terminated in diabase that is overlain by undifferentiated Lower Cretaceous or older Mesozoic clastic rocks. Three wells in north Florida penetrated weathered(?) diabase that underlies Lower Cretaceous clastic rocks and rests on Early Paleozoic strata. Two other wells in the northeast part of the Florida peninsula and a well in southeast Georgia penetrated diabase sills and dikes that cut early Paleozoic strata. A single well in Jackson County, Florida, situated west of the Peninsular Arch, also penetrated a small intrusive or flow of porphyritic hornblende basalt in lower Paleozoic sedimentary rocks.

The time of emplacement of the diabase in the subsurface formations in the Southeastern States has not been precisely defined, but on

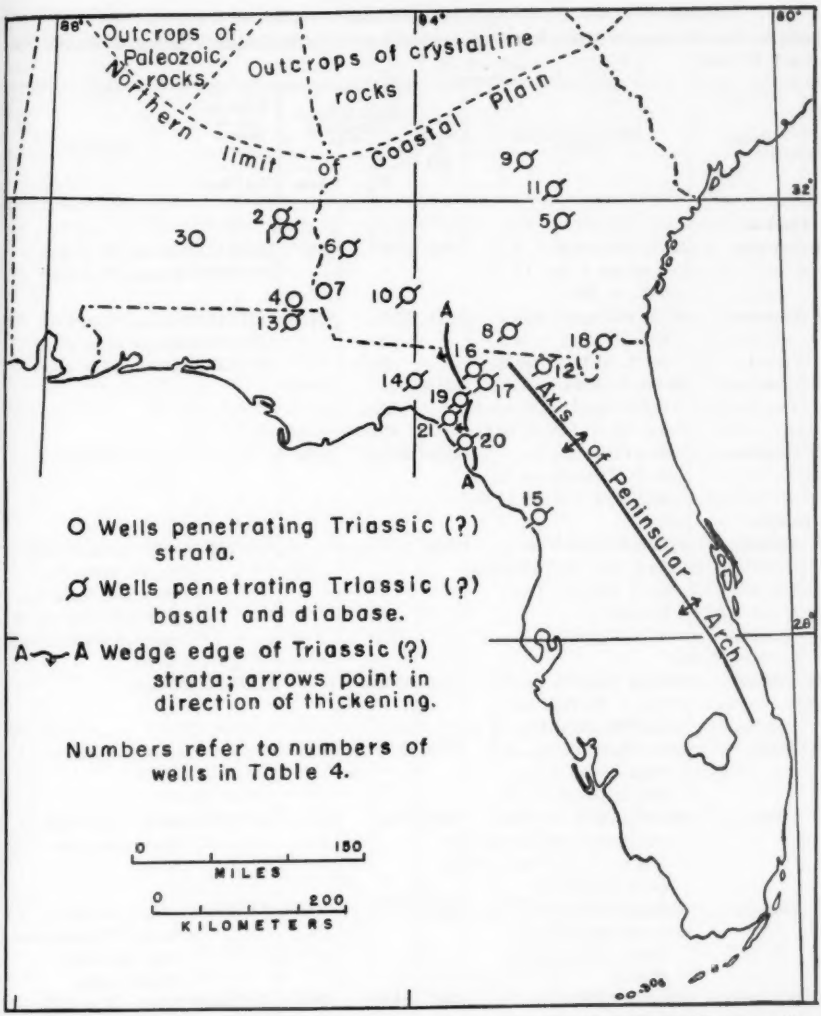
Outcrops of Paleozoic rocks
Outcrops of crystalline rocks
Northern limit of Coastal Plain
Axis of Peninsular Arch
O Wells penetrating Triassic (?) strata.
Ø Wells penetrating Triassic (?) basalt and diabase.
A—A Wedge edge of Triassic (?) strata; arrows point in direction of thickening.
Numbers refer to numbers of wells in Table 4.
0 150
MILES
0 200
KILOMETERS

FIGURE 6.—MAP OF FLORIDA AND PARTS OF ADJACENT STATES SHOWING LOCATION OF WELLS IN WHICH TRIASSIC (?) ROCKS HAVE BEEN FOUND AND WHICH ARE LISTED IN TABLE 4

the basis of available data its age is tentatively fixed as Late Triassic. The pre-Cretaceous age of the diabase is emphasized by the lack of indications of volcanism in the Cretaceous and younger formations in the southeastern region (Applin, 1951, p. 16). Concerning the diabase penetrated in a well in Nassau County in extreme northeastern Florida, J. O. Fuller (Cole, 1944, p. 94) stated "The diabase of the Florida sample agrees surprisingly well in mineral composition with the upper chilled zone of the Palisade diabase. The sample from the Florida well is tentatively classified as Triassic." According to Moody (1949, p. 1410, 1411), certain diabase dikes which occur in the Piedmont province in southwest Georgia have been referred to the Triassic(?), and one of them passes unconformably beneath the Coastal

TABLE 4.—TRIASSIC (?) ROCKS IN WELLS IN ALABAMA, GEORGIA, AND FLORIDA

Location of wells shown in Figure 6. Figures marked * are from Southeastern Geological Society cross-section E-E′ (1941).

Well no. on Fig. 6	County	Name and location of well	Total depth of well	Depth in feet to sedimentary rocks		Depth in feet to igneous rocks		Remarks
				Top	Base	Top	Base	
	Alabama							
1	Barbour	W. B. Hinton *et al.*, J. S. Creel no. 1. Sec. 14, T. 9 N., R. 26 E.	5546	4400*	5546	5342 5491	5372 5522	Diabase sill or dike Diabase sill or dike
2	Barbour	H. A. Stebinger, Alice S. Robertson no. 1. Sec. 19, T. 10 N., R. 26 E.	5215	Not identified	5215	4135 4202 4273	4152 4208 4274	Diabase sill or dike Diabase sill or dike Diabase sill or dike
3	Crenshaw	Nelson Exploration Co., Smith Lumber Co. no. 1. Sec. 26, T. 8 N.,R.16 E.	10830	Not identified	10830			
4	Houston	Union Producing Co., E. P. Kirkland no. 1. Sec. 20, T. 7 N.,R. 11 W.	8100	6990±	7556			Lower Paleozoic at 7556
	Georgia							
5	Appling	Felsenthal and Weatherford, Mrs. W. E. Bradley no. 1. Lot 522, Land District 2	4106			4095	4106	Altered amygdaloidal basalt, probably a lava flow; underlies Lower Cretaceous or older Mesozoic clastic rocks
6	Calhoun	Sowega Minerals Explor. Co., J. W. West no. 1. Lot 328, Land Dist. 4	5265	3850±	5190	5190	5265	Diabase
7	Early	Mont Warren et al., A.C. Chandler no. 1. Lot 406, Land district 26	7320	5670*	6950			Lower Paleozoic at 6950
8	Echols	Humble Oil & Refining Co.,Bennett and Langsdale no. 1. Lot 146, Land District 12	4185			4125	4150	Diabase sill or dike in lower Paleozoic strata
9	Laurens	Calaphor Manufact'g Co., Grace McCain no. 1. Half a mile south of Minter	2548			2532	2548	Diabase, underlies Lower Cretaceous or older Mesozoic clastic rocks
10	Mitchell	Stanolind Oil & Gas Co., J. H. Pullen no. 1. Lot 133, Land District 10	7487	6400±	7487	6550 7070 7350	6620 7090 7470	Diabase sill or dike Diabase sill or dike Diabase sill or dike
11	Montgomery	J. E. Weatherford, Lonnie Wilkes no. 1. Half a mile south of Higgston	3433			3415	3433	Diabase, underlies Lower Cretaceous or older Mesozoic clastic rocks
	Florida							
12	Columbia	Humble Oil & Refining Co., J. P. Cone no. 1. Sec.22,T. 1 N.,R.17 E.	4444			3529 3564 4191 4193 4248 4267	3562 3565 4192 4195 4251 4270	Diabase sills or dikes in early Paleozoic strata

TABLE 4.—*Continued*

Well no. on Fig. 6	County	Name and location of well	Total depth of well	Depth in feet to sedimentary rocks		Depth in feet to igneous rocks		Remarks
				Top	Base	Top	Base	
	Florida Cont'd							
13	Jackson	Humble Oil & Refining Co., C. W. Tindle no. 1. Sec.8,T. 5 N.,R. 11 W.	9245	8220±	8440	8890 8970	8932 8983	Basalt intrusives or flows in early Paleozoic strata;Triassic(?) clastics to 8440
14	Jefferson	Coastal Petroleum Co., E. P. Larsh no. 1. Sec. 1, T. 2 S., R. 3 E.	7913	7030±	7909?	7763 7850	7792 7890	Diabase sill or dyke Diabase sill or dike Triassic(?) clastics to 7909
15	Levy	Humble Oil & Refining Co., C. E. Robinson no. 1, Sec.19,T.16 S.,R.17 E.	4609			4317	4377	Decomposed igneous rock to 4344; underlies Lower Cretaceous. Basalt to 4377; overlies early Paleozoic
16	Madison	Hunt Oil Co., J. W. Gibson no. 2. Sec. 6, T. 1 S., R. 10 E.	5385			4589	4628	Diabase; underlies Lower Cretaceous clastic rocks, overlies early Paleozoic strata
17	Madison	Hunt Oil Co., J. W. Gibson no. 4. Sec. 5, T. 2 S., R. 11 E.	4096			4044	4060	Diabase; underlies Lower Cretaceous clastic rocks, overlies early Paleozoic strata
18	Nassau	St. Marys River Oil Corp., Hilliard Turpentine no. 1. Sec. 19, T. 4 N., R. 24 E.	4824			4808	4824	Diabase sill or dike; underlies early Paleozoic strata
19	Taylor	Gulf Oil Corp., Brooks-Scanlon, Inc., Block 33, no. 1. Sec. 18, T. 4 S., R. 9 E.	5243	Not identified	5200	5200	5243	Diabase gabbro
20	Taylor	Gulf Oil Corp., Brooks-Scanlon, Inc., Block 42 no. 1. Sec. 9, T. 8 S., R. 9 E.	5517	5120±	5438	5438	5517	Diabase, probably a flow
21	Taylor	Humble Oil & Refining Co., G.H.Hodges no.1. Sec.12, T. 5 S.,R. 6 E.	6254	5960±	6153	6153	6254	Basaltic rock to 6165, Diabase gabbro to 6254

Plain deposits. Moody (1949, p. 1411–12) stated further "The age of these dikes has been tentatively fixed as Triassic(?) because diabase is common as a Triassic intrusive and because field evidence establishes the time of emplacement to be in the interval between the close of Appalachian orogeny and the beginning of Upper Cretaceous time." Lester and Allen (1950, p. 1217), on the contrary, state that no definite age can be given to the diabase in the Georgia Piedmont. (*Applin*)

63. *Columns 75, 76. North Carolina.*—Triassic sedimentary rocks are exposed in two northeast-trending belts in the Piedmont plateau of North Carolina. The easternmost belt includes the Deep River and Wadesboro Triassic basins, troughlike structural and topographic basins separated by an overlap of

unconsolidated Pliocene(?) sand and gravel along the inner edge of the Coastal Plain. The westernmost belt includes the Dan River Triassic basin and a small outlier of Triassic rocks southwest of this basin. Rocks in both belts were assigned to the Newark Group by Russell (1892).

In the Deep River and Wadesboro basins, the Triassic sedimentary rocks, 8000 to 10,000 feet thick, were grouped into three formations by Campbell and Kimball (1923) and were further defined and described in detail by Reinemund (1955). The Pekin Formation at the base of the Triassic sequence lies unconformably on Paleozoic and Precambrian igneous and metamorphic rocks; it is composed of brown, red, and gray claystone, shale, siltstone, sandstone, and conglomerate. This unit is overlain by the Cumnock Formation containing gray and black shale, siltstone, and sandstone, and two beds of coal. At the top of the Triassic sequence is the Sanford Formation, composed of brown, red, and gray claystone, shale, siltstone, sandstone, conglomerate, and fanglomerate.

In some parts of the Deep River and Wadesboro basins the Cumnock Formation grades laterally into brown and red rocks indistinguishable from strata in the Pekin and Sanford Formations. In these areas, the strata correlative with the Cumnock are included in the Pekin and Sanford, and the contact between the Pekin and the overlying Sanford is obscure and gradational.

Studies of fossil wood from the base of the Pekin Formation by Knowlton (*in* Ward, 1900, p. 270–274) and of fossil plants from the Cumnock Formation by Fontaine (1883) have established that these rocks are of Late Triassic age.

In the Dan River Basin, the Triassic sedimentary rocks consist of three units, probably totaling about 8000 feet in thickness, which are lithologically similar to the Pekin, Cumnock, and Sanford Formations in the Deep River and Wadesboro basins. Although these units have not been named, they have been described by Stone (1910), who regarded them as correlative with the rocks in the Deep River Basin and in the Richmond Basin of Virginia. The middle unit is thinner and contains less valuable beds of coal than the Cumnock Formation but occupies approximately the same stratigraphic position. (*Reinemund*)

64. *Columns 77, 78. Virginia.*—Triassic sedimentary rocks in Virginia are exposed in the Potomac, Scottsville, Danville, Farmville, and Richmond basins and in several small outliers of these basins in the Piedmont plateau. In all except the Richmond basin, the Triassic sequence consists mainly of brown and red strata (red beds) similar to the Triassic rocks in other Atlantic Coastal States. In the Richmond basin, the Triassic rocks are mainly gray and black and contain several coal beds.

The Triassic sequence in the Richmond Basin was divided by Shaler and Woodworth (1899) into the Boscabel Boulder Beds, Lower Barren Beds, Productive Coal Measures, Vinita Beds, and Otterdale Sandstones. They included the Vinita Beds and Otterdale Sandstones in their Chesterfield Group, and the lower units in their Tuckahoe Group, but inasmuch as the contemporaneity of the coal-bearing rocks in the Richmond Basin and in the North Carolina basins was established by Fontaine (1883), and the name Newark Group was subsequently applied by Russell (1892) to all Triassic rocks in Virginia and North Carolina, it seems best to retain the name Newark Group for the Triassic of the Richmond Basin.

In a general study of all Triassic basins in Virginia, Roberts (1928) applied the names Border Conglomerate, Manassas Sandstone, and Bull Run Shale to rocks in all the basins, including the Richmond basin. Inasmuch as the rocks to which these names are applied in other basins are lithologically different from the rocks in the Richmond basin, the earlier names used by Shaler and Woodworth have been retained in the chart, and the Richmond basin has been treated separately.

The Triassic sequence in the Richmond basin was estimated by Shaler and Woodworth to be somewhat more than 3300 feet thick, but this estimate is in doubt because undiscovered faults probably exist within the basin. Although this sequence appears to be less than half as thick as the rocks in other Triassic basins in Virginia and North Carolina, it may represent nearly as much time because the coal and shale that constitute a large part of the sequence in the Richmond basin probably accumulated more slowly than the coarse clastic rocks that constitute a large part of the sequence in other basins. If the coal beds near the base of the Triassic sequence in the Richmond basin are approximate time equivalents of the coal beds in other basins, rocks equivalent to the basal part of the Newark Group in the other basins may be missing, as indicated on the chart.

Triassic rocks in the Potomac, Scottsville, Danville, and Farmville basins consist mainly of brown or red claystone, shale, siltstone, sandstone, and conglomerate. The Farmville basin also contains several thin beds of coal in a unit of gray and black shale and sandstone. The Danville basin contains a similar lenticular unit of dark shale and sandstone.

Roberts (1928) applied the names Border Conglomerate, Manassas Sandstone, and Bull Run Shale to rocks of corresponding lithology in the different Triassic basins in Virginia. Although he regarded the conglomerate as generally the oldest rock and the shale as generally the youngest rock in the Triassic sequence, the cross sections accompanying his report show that he applied these names to units of similar lithology in all parts of the stratigraphic sequence. Thus, the names applied by Roberts have lithologic rather than time-stratigraphic significance and are not directly comparable to the terminology in other basins.

The lenticular units of dark shale and sandstone in the Farmville and Danville basins have not been named, and their stratigraphic position has not been precisely determined.

(*Reinemund*)

65. *Columns 79–92.—Sedimentation and correlation of Newark Group.*—The sediments of the Newark Group are continental and appear to have been laid down in a number of long, narrow intermontane troughs that trended nearly parallel to the Appalachian folded structures, although they cut across them locally at acute angles. The hypothesis that Newark sedimentation was initiated by faulting which formed grabens is not supported by the distribution of faults or by the character of the earliest sediments. Thus, in the Pennsylvania area, faults occur along parts of the northwest border of the basin, whereas the lowest formation was clearly derived from a southern provenance. No border faults have been found along the southern margin. A hypothesis (McLaughlin, 1953) that accords well with the facts attributes the Newark basins to deep erosion of the least resistant of the folded Paleozoic rocks, while the crystalline Piedmont remained as a lofty highland to the south of the largest known Triassic trough (the New York-Virginia belt). Along its northwestern side a lesser escarpment of Silurian quartzite formed the limit of the basin.

The first sediments originated mainly by erosion of the crystalline massif to the south (McLaughlin and Gerhard, 1953) and are arkosic: the Stockton and New Oxford formations. Similar rocks form the lowest member of the Newark throughout its known range. After the highland had been considerably lowered by erosion, drainage in the basin became sluggish, and a broad, shallow lake with marshy borders occupied the center of the trough in its widest portion (eastern Pennsylvania and western New Jersey). Here the black shale of the Lockatong Formation accumulated, and in these beds are found the most abundant fossils. Later brief recurrence of lacustrine or swampy conditions is indicated by black shale members well up in the next overlying red shale formation.

In the meantime, faulting occurred in many places along the northwestern escarpment, and the amount of sediment washed in from that side increased. The formations of Paleozoic age furnished much quartzose sand and silt. Long-accumulated red regolith, supplemented by red detritus from Paleozoic red beds, accounts for the great development of red shales and sandstones in the upper formations of the Newark Group: the Brunswick in New Jersey and eastern Pennsylvania, and the Gettysburg farther west and south. Where the basins were narrower, lakes and swamps had only a brief existence or did not develop at all, and the arkosic sedimentation was supplanted by deposition of quartzose sands and silts, and red muds from the north. Great alluvial fans formed along the northwestern margin of the basin (McLaughlin, 1939), and some of these extended far out toward the center. Near the center of the trough, the southern arkosic and the northern quartzose facies interfingered (McLaughlin and Willard, 1949). Where the lacustrine deposits developed, they interfingered with the red shale and sandstone formation.

The strong development of fanglomerates along the present northwest border of the Triassic shows that the limit of sedimentation in that direction was close to the present location of the border. Although faults are known at some places, at others the Triassic rocks appear to overlap the older rocks (Wherry, 1941; McLaughlin, 1946), as if piled against a steep escarpment. Faulting may have occurred at all these places, but, if so, the Triassic sediments have buried the fault, and there has been no large post-Newark movement. Probably the Triassic sediments never extended more than 2–3 miles north of their present limit except in channels of the streams

that flowed into the basin on that side. The Spitzenberg Conglomerate (Whitcomb, 1942), if it is Triassic, may be such a stream-channel deposit. It seems impossible that it could be a remnant of a great sheet of sediments once continuous with the main Triassic belt some 18 miles to the south.

Thicknesses of the series are many thousands of feet. Sections in the Gettysburg and Delaware River areas are typical.

Gettysburg area (Stose, 1929)
- Gettysburg Formation, 16000 feet
- New Oxford Formation, 7000

Delaware Valley (Pa. Topog. Geol. Survey Rept. Bucks County)
- Brunswick Formation, above Haycock still, 5000 feet
- Brunswick Formation, below Haycock sill, 3800 feet
- Lockatong Formation, 3800 feet
- Stockton Formation, 5000 feet

It is not to be expected that the full thickness of all the formations would be present at any one place. The younger beds overlap the older northward and westward, and the oldest probably thin out north of the center of the basin. They may disappear near the north border. Evidence of steep shelving of the floor on that side has been found. Probably the thickness present in any vertical section does not exceed two-thirds of the total. It has sometimes been suggested that an illusory great thickness has been produced through repetition by numerous strike faults. This hypothesis is disposed of by demonstrated lack of faulting in some thick exposed sections, and by tracing of identifiable beds along their curved strike in gently plunging synclines. The thickness of the series is indeed great.

Fossils are scarce in the Newark Group, but some localities have yielded fairly numerous specimens. None of the fossils has yet furnished a precise dating of any bed. A large amphibian was found near Holicong, Bucks County, Pennsylvania, near the base of the Stockton (Sinclair, 1917). At Fort Lee, New Jersey, a specimen of *Clepsysaurus* was taken from beds below the Palisade diabase (Von Huene, 1913). Somewhat above the middle of the New Oxford in York County, Pennsylvania, numerous fresh water mollusks were found (Wanner, 1921; 1926). Similar mollusks were collected near the top of the Stockton at Montclare, Montgomery County, just across the Schuylkill from Phoenixville (Richards, 1948). In York County, 31 species of plants were described from the upper beds of the New Oxford (Wanner and Fontaine *in* Ward, 1900, p. 233–254). A quarry near the top of the Stockton at Carversville, Bucks County, has yielded numerous plants (Brown, 1911; Wherry, 1916; Bock, 1952a). Fish and estherids from black shale at North Bergen, New Jersey (Schaeffer, 1941), have been considered of late Stockton age but the exposure may be a small thinned extension of the Lockatong.

Localities in the Lockatong Formation: Many years ago (Cope, 1866) the Phoenixville tunnel was a source of various reptilian remains in the Lockatong Formation. A deep railroad cut at Gwynedd, Montgomery County, Pennsylvania, has been a rich hunting ground. Fish, estherids, a small reptile, and an amphibian have been described (Bryant, 1934; Bock, 1945; 1946; Von Huene and Bock, 1954). A jaw of *Clepsysaurus* was found at Belfry, Montgomery County (Colbert, 1943). At Princeton, New Jersey, numerous fish were collected from the excavation for the University library (Jepsen, 1949).

Colbert (1946) has summarized most of the information on Brunswick fossils in New Jersey and their correlation with those of the Connecticut Valley. Especially noteworthy are the numerous fish from Boonton, New Jersey (Eastman, 1905), above the highest lava flow. Reptile skeletons (*Hypsognathus*) have been found at Clifton and Passaic, below the lavas (Gilmore, 1928; Colbert, 1946). Dinosaur footprints and plants were collected and described from Milford, New Jersey (Hitchcock, 1889; Newberry, 1888c; Bock, 1952a; 1952b). The type specimen of *Clepsysaurus* was found near Hosensack, Lehigh County, Pennsylvania (Lea, 1853). Numerous dinosaur footprints occur in the upper beds of the Gettysburg shale in York County (Hickock and Willard, 1933; Willard, 1934).

The correlation of the eastern Triassic with the standard European sections has been attempted for a century or so, with conflicting results. Probably few would now dissent from the opinion that the Newark is practically all Upper Triassic, but any attempt at closer definition would bring out considerable differences of opinion.

Concerning the Newark as a whole, we cite a few of the suggested correlations. Ward (1892), on the basis of the plants, considered these beds to represent the "summit of the Triassic", the upper Keuper or Rhaetic.

Eastman (1905) studied the fish from Boonton, New Jersey, near the top of the series and above the lava flows. He concluded that the fauna was a local one and insufficient for precise correlation, but he pointed out similarities to forms from the Virglorian Muschelkalk (Middle Triassic) and recognized some admixture from the Keuper (Upper Triassic). Later, Eastman (1911) studied the fish of the Connecticut Valley, chiefly from between the lava flows somewhat above the middle of the sedimentary series. He stated that this fauna was more or less composite and contained both Ladinian (Middle Triassic) and Keuper forms. He further indicated that the most satisfactory correlation of the upper unit was with the Raibl beds of the lower Keuper.

Wherry (1912) appears to have considered that the Newark occupied a long period in the Middle and Upper Triassic. His views were similar to those later advanced by von Huene.

Lull (1915, p. 20), from a study of the land vertebrates, chiefly from the Connecticut Valley, placed the Newark considerably higher than did Eastman. Though he admitted the dating was not precise, he pointed out that the dinosaur remains from the sandstones below the lava flows were most nearly allied to middle Keuper forms, whereas the abundant footprints and some skeletal remains in the beds above the lavas all indicated an age not earlier than late Keuper. He even suggested that the Newark might extend through the time between Triassic and Jurassic. In a revision, Lull (1953, p. 17) apparently abandoned this last suggestion but reaffirmed the middle to late Keuper age.

Berry (1923) called the Newark "not older than the Keuper . . . nor younger than the uppermost Triassic".

Von Huene (1926) favored a more extended period for the Newark. Its end was considered coincident with the end of Triassic time; its beginning was placed well down in the Muschelkalk. Thus, as regards the evidence from the land vertebrates (from the middle to upper beds of the sequence) he was not in disagreement with the view later adopted by Lull (1953) but he placed the beginning considerably earlier.

Colbert (1946, p. 266) summarized the paleontological evidence and favored a Late Triassic age, although he admitted that Von Huene's suggestion of a beginning in the Muschelkalk was tenable.

Recently, Bock (1952b) suggested a closer correlation and the shortest period of time yet proposed. According to him, the Newark corresponds almost entirely to the middle Keuper, with the youngest beds perhaps extending into the lower part of the Rhaetic. However, the available evidence seems hardly to justify such a narrow restriction, since most of the fossils have too large a range in time for precise dating. The great thickness of sediments appears to require a long period, although these continental deposits must have been laid down rather rapidly. Probably the period proposed by Bock is not impossibly short.

The question of exact dating of the Newark Group still appears to be open; the extreme views are those of Von Huene (1926) and of Bock (1952b). For the present it seems best to adopt a somewhat arbitrary compromise view that the Newark is roughly the equivalent of the greater part of the Keuper, or Upper Triassic.

Correlation between separated belts of Newark rocks is in no better state. In all probability the deposition in the various areas was roughly contemporaneous, but there is no proof of equivalence of specific beds from region to region. There are, at best, only some suggestive correspondences.

On paleontological grounds, Lull (1915, p. 80; 1953, p. 62) apparently considered it possible that the Connecticut Valley sediments were in part later than those farther south. Thus:

> "It is my opinion that the Pennsylvania localities represent an older faunal phase than the dinosaurs of the Connecticut Valley."

This view is consistent with the suggestion of Schuchert and Dunbar (1941) that the three lava flows in New Jersey were synchronous with the three flows in Connecticut. This equivalence was also proposed by Colbert (1946, p. 266), who pointed out that the distribution of the various fossils in the two areas and of the footprints in the upper shales of the Connecticut Valley supported it. Jepsen (1948) appears to concur in a tentative way. Bock (1952b), also on paleontological grounds, considers that sedimentation began later in the Connecticut Valley than in the Delaware Valley. He says:

> "It should be concluded that the Triassic of the Connecticut Valley comprises the facies of the Stubensandstein to the Knollermergel [upper middle Keuper] with overlap in the central portion of the middle Keuper, but that the lower facies of the middle Keuper [present in Pennsyl-

vania] are wanting." (Bock, 1952b, p. 430; the brackets are by McLaughlin).

Bock also claimed that the deposits of the Richmond basin are all later than the Lockatong of the Delaware River section, that is, all upper middle Keuper. He also regarded the coal-bearing Cumnock Formation of the Deep River area of North Carolina as the equivalent of the Lockatong.

Any attempt to correlate between separate belts by means of lithologic similarities of stratigraphic sequences would entail grave risks. It is reasonable to expect that the extensive igneous activity in New Jersey and Connecticut would have been simultaneous, since the actual separation of the two areas only slightly exceeds the extent of the flows in New Jersey and is much less than that of the lavas in Connecticut. The correlation based on equivalence of the three flows therefore appears fairly probable. On the other hand, the commonly observed division of the Newark Group into a lower arkose and an upper red shale and sandstone unit has probably no real correlational value. Where granitic rocks were exposed under similar topographic and climatic conditions, the first sediments would be arkose, whether the deposition began early or late. The sequence of three formations in the Deep River basin of North Carolina resembles that in the Delaware River section, and some workers have assumed equivalence. The similarity, however, may be circumstantial. Bock considers the Cumnock equivalent to the Lockatong, but none of the fossils furnishes conclusive dating.

The situation is more encouraging in the largest single belt, which extends almost uninterruptedly from New York to northern Virginia. In this region, correlation can be carried from area to area by stratigraphic continuity. Even this method can be objected to, on the ground that undiscovered faults may exist, but this objection loses much of its force as stratigraphic and structural details become better known. Actually, the greatest source of uncertainty is not the possibility of unknown faults, but the probability that interfingering of facies causes certain contacts to rise or descend stratigraphically.

Several recognizable thick members persist for many miles along the strike and can be used as key beds. Near the northern extremity of the belt, the Watchung lavas serve as marker beds high in the Brunswick Formation. Several thousand feet lower, the black shale and argillite facies of the Lockatong Formation is strongly developed in west-central New Jersey and continues many miles into Pennsylvania. Though precise equivalence cannot be established for individual beds throughout its range, the base of the Lockatong is a well-defined horizon, and interbedded red shales above the middle have been traced for a number of miles (Pa. Topog. Geol. Survey, 1950). Still higher, the black shale of the Graters and Perkasie Members in the Brunswick are reliable key beds from the Delaware nearly to the Schuylkill (McLaughlin, 1948).

The Lockatong disappears in Chester County, Pennsylvania, and the higher key beds are lost also. The entire section above the Stockton arkose grades westward into sandstone and conglomerate. Within the thick Robeson Conglomerate, however, interbedded red shale members and some ridge-making conglomerates define the strike well enough to permit the conclusion that the topmost beds of the Stockton in the east and of the New Oxford to the west are not far apart stratigraphically.

A large cross fault in Lancaster County interrupts the continuity, but the New Oxford and Gettysburg contact is well defined west of it. In the Gettysburg Formation a number of prominent conglomerate members have been traced for miles (Pa. Topog. Geol. Survey, unpublished data; McLaughlin and Gerhard, 1953). With the aid of these and of the New Oxford and Gettysburg contact, correlation is carried westward beyond the Susquehanna. Locally the New Oxford Formation seems to rise higher stratigraphically, interfingering with the lower Gettysburg, but, from western York County southward through Maryland, this contact conforms closely to the strike of the beds. In York and Adams counties the thick Heidlersburg Member is a well-defined zone. In Maryland considerable faulting is evident, but the New Oxford-Gettysburg contact is readily recognizable. (*McLaughlin*)

66. *Column 79. Montgomery county, Maryland, and northern Loudoun county, Virginia.*—On the map of Montgomery County (Maryland Dept. Mines, Geology and Water Res., 1953) the Triassic is undifferentiated and is indicated as "largely New Oxford." The following interpretation is based chiefly upon a reconnaissance by the compiler (McLaughlin, unpublished).

Both New Oxford and Gettysburg formations are present in Montgomery County. A fault with northeast strike, shown in part on

the county map, offsets the base of the Triassic, as well as the New Oxford and Gettysburg contact, and causes repetition of several thousand feet of strata. The fault continues at least to the Potomac River, south of which it has not been traced, as it must lie wholly in the area of shale of the Gettysburg (Bull Run). A second major fault cuts off the Triassic on the west at Monocacy River, where the Potomac bends sharply southwestward. This fault apparently continues southwestward along the Potomac beyond Ball's Bluff. West of this fault, in northern Loudoun County, Virginia, the entire thickness of the New Oxford is again exposed, directly on the strike of that formation in southern Frederick County.

Immediately above this western band of New Oxford is a great thickness of interbedded limestone fanglomerate and red shale (Bull Run). These rocks were mapped accurately by Keith (1894). The western border of the Triassic here is probably determined by a large fault. The development of limestone fanglomerate at this locality is the greatest known in the Triassic areas. (*McLaughlin*)

67. *Column 80. Frederick and Carroll counties, Maryland.*—The area was mapped and described by Stose and Stose (1946). Near Frederick, the pre-Triassic floor is locally exposed entirely across the Triassic belt. Along the margin of the Triassic area at this interruption, a continuous narrow band of limestone conglomerate represents the base of the New Oxford, but its trend is transverse to the strike of the westward-dipping beds. The compiler interprets it as a "plaster conglomerate" on the sloping limestone floor upon which the Triassic was deposited. Along the western margin of the Triassic in southern Frederick County is a wide band of limestone fanglomerate, continuous with that in northern Loudoun County Virginia.

In northeastern Frederick and northwestern Carroll counties, the normal sequence of arkose and sandstone of the New Oxford Formation, overlain by Gettysburg shale, occurs in great thickness and is continuous along the strike with the formations in the Gettysburg district, Adams County, Pennsylvania. In Maryland the contact of New Oxford and Gettysburg appears to be a strike line, with little indication of interfingering along the contact. (*McLaughlin*)

68. *Column 81. Eastern shore (Wicomico and Worcester counties), Maryland.*—Rocks assigned on lithologic composition to the Newark Group are reported in two wells on the Eastern Shore of Maryland (Richards, 1945; Anderson and others, 1948, p. 13, 83, 100). In the Hammond well, 6 miles east of Salisbury, at a depth of 5360 feet, 135 feet of hard quartz conglomerate underlain by reddish-brown and apple-green shales, sandy shales, and arkosic sandstones was noted. In the Bethards well, about 17 miles southeast of Salisbury, 585 feet of silty reddish-brown shale, mottled reddish-brown and bottle-green sandstones interbedded with chocolate-brown shales and occasional lead-gray strata, and a basal congolomerate was noted. (*McLaughlin*)

69. *Column 82. Adams county, Pennsylvania.*—The New Oxford Formation here attains its maximum thickness and is overlain conformably by Gettysburg shale. Above the middle of the Gettysburg section is the Heidlersburg Sandstone Member. Near the top of the section, fanglomerate occupies considerable areas along the northwest border. The Heidlersburg Member grades along the strike into a part of the fanglomerate, which in turn lies unconformably on Ordovician limestone. Near Bendersville, a small basalt flow occurs in the fanglomerate (Stose, 1929). (*McLaughlin*)

70. *Column 83. York county, Pennsylvania.*—The New Oxford Formation is extremely thick, but apparently its uppermost beds interfinger with the lower part of the Gettysburg and grade into the latter northeastward along the strike. Above the base of the Gettysburg, the thick lens of Conowago Conglomerate fingers out both northeastward and southwestward into normal Gettysburg. The Heidlersburg Member continues nearly to the Susquehanna River, where diabase cuts across it. Fanglomerates are developed at a few places near the border. At most places the normal red shales near the north border contain thin interbeds of limestone-and-shale conglomerate, derived from the neighboring Paleozoic (Stose and Jonas, 1939).

In a new exposure on the Pennsylvania Turnpike, red calcareous shale and conglomerate unconformably overlie limestone of Paleozoic age. (Pa. Topog. Geol. Survey, unpublished). (*McLaughlin*)

71. *Column 84. Dauphin and western Lancaster counties, Pennsylvania.*—Here the New Oxford and Gettysburg Formations are relatively uncomplicated. Conglomerates are developed in the uppermost part of the Gettysburg within a few miles of the north border. (Stose and Jonas, 1933). (*McLaughlin*)

72. *Column 85. Lebanon and Lancaster counties, Pennsylvania.*—The New Oxford is very thick in western and northern Lancaster County, but it thins rapidly near the eastern County line (Jonas and Stose, 1926; 1930). It is quite certain, in view of the evidence from Berks and Chester counties, that this change represents a shelving of the floor of the depositional basin toward the east. In the northern part of Lancaster County, the New Oxford interfingers with the lower beds of the Gettysburg and grades eastward along the strike into sandstone and then into the lower beds of the Elizabeth Furnace Conglomerate Lens (McLaughlin and Gerhard, 1953). Several thinner conglomerate lenses occur higher in the Gettysburg, which is here mainly sandstone. At the north border, fanglomerate occurs locally, and there is evidence that it lies directly upon pre-Triassic rocks but interfingers with sandstones and shales high in the Gettysburg (Pa. Topog. Geol. Survey, unpublished material). (*McLaughlin*)

73. *Column 86. Berks and Chester counties, Pennsylvania.*—Here the Stockton is the equivalent of the New Oxford in areas to the west. The Stockton thickens rapidly eastward from about 1000 feet near Coventryville to more than 2000 feet near Phoenixville. This is quite definitely due to a deepening of the basin. The Stockton Arkose, followed westward along the strike, ends against the Precambrian rocks of the Honeybrook upland, which must have been undergoing erosion during at least the earlier two-thirds of Stockton time.

In eastern Chester County, a thin member of black and greenish shales first appears. This Lockatong Formation is a local facies, the time equivalent of the lower part of the Brunswick. It thickens eastward from zero to about 1500 feet near Phoenixville. It interfingers with red shale and sandstone of the Brunswick.

In southern Berks County, the position of the Gettysburg Formation is occupied by a thick sequence of conglomerate and sandstone, the Robeson Conglomerate. Westward it fingers out into the Gettysburg; eastward it interfingers with the red shale and sandstone of the Brunswick Formation, which is the eastern lithologic and stratigraphic equivalent of the Gettysburg. (Bascom, 1938; McLaughlin, 1939).

At the north border, extensive limestone fanglomerate interfingers with the upper shale beds of the Brunswick, and this fanglomerate in turn lies unconformably on (or against) the pre-Triassic limestones. Near the top of the series a flow of vesicular basalt is closely associated with shales and fanglomerate (Wherry, 1910). Although part of the border is marked by faults, at several localities the fanglomerate and other Triassic sediments are unconformable on the older rocks. The Newark Group apparently accumulated against a steep escarpment. Relations observed many years ago in the Boyertown iron mines (d'Invilliers, 1883) make it plain that the Triassic is not simply faulted down against the limestone, but overlies it unconformably. (*McLaughlin*)

74. *Column 87. Montgomery and Lehigh counties, Pennsylvania.*—The Newark rocks in Montgomery County were mapped by Darton (1909) and Wherry (1931). Interpretation in following paragraphs is based in part on unpublished work by the compiler. (*See* also McLaughlin, 1939; 1948.)

The Stockton Arkose continues to thicken eastward to more than 4000 feet near Norristown. Indicative of the steepness and proximity of the border of the basin is the coarseness of some of the arkose at Norristown.

The Lockatong Formation thickens from 1500 feet at the Schuylkill to nearly 3000 feet near the eastern Montgomery County line. This change is due mainly to the appearance, in the lower beds of the Brunswick, of thin beds of the Lockatong, which thicken eastward by gradation from red shale of the Brunswick to black or gray shale and argillite of the Lockatong.

The Brunswick Formation is mainly red shale, with interbedded sandstones. In Lehigh County, fanglomerate interfingers with the upper part of the Brunswick and lies upon Paleozoic limestone. Here again is indication of a steep shelving floor or an escarpment against which the Triassic sediments accumulated (Wherry, 1941).

The Graters Member consists of thick dark-gray shale submembers, separated by red shale. It persists for many miles along the strike and is the clearest horizon marker in the entire region (McLaughlin, 1948).

The Perkasie is a dark shale member as conspicuous as the Graters. For several miles it lies so close to a belt of metamorphosed shale below a thick diabase sill that it was long mistaken for part of the baked shale. However, it is well separated in western Montgomery County and is recognizable throughout its outcrop (McLaughlin, unpublished material).

(*McLaughlin*)

75. *Column 88. Bucks county, Pennsylvania.*—The Triassic rocks in Bucks County have been mapped in great detail (Pa. Topog. Geol. Survey, 1950, and unpublished report; *see* also McLaughlin, 1943; 1944; 1945).

The Stockton Formation reaches its maximum thickness in the northern of two belts of outcrop, in central Bucks County. Here, a section of 5000 feet has been measured. A prominent quartz conglomerate member occurs several hundred feet above the base.

The Lockatong Formation also is at its maximum development in the northern belt of outcrop, with a thickness of 3800 feet. In this section the extensive interfingering of Lockatong and Brunswick is exhibited best. Above the main body of the Lockatong, a thickness of more than 1000 feet consists of alternating thick red and black shale units. The black shales all thin westward and grade in that direction into red shale.

The Brunswick Formation here is essentially red shale. Only near the north border is there any considerable amount of sandstone. The Graters and Perkasie Members continue across the county with very little change.

At the north border, limestone and quartz fanglomerate interfinger with red shales high in the Brunswick. The north border is in part determined by a major fault, but there is evidence that some of the fanglomerate overlapped and buried the fault. (*McLaughlin*)

76. *Column 89. Hunterdon county, New Jersey.*—This area includes the type localities of the Stockton and Lockatong (Kummel, 1898). The formations are similar in character to those of Bucks County, Pennsylvania, with which they are stratigraphically continuous.

Along the north border is the most clearly demonstrated example of shelving of the floor of the depositional basin. The Stockton, Lockatong, and Brunswick Formations pass northward along the strike, through a gradational sandstone with interbedded conglomerate, into a large area of fanglomerate that constitutes the border facies of all three formations. This conglomerate in turn lies unconformably upon Cambrian and Ordovician rocks. Apparently these rocks represent part of a rather steep escarpment, at whose foot the alluvial gravels accumulated, while the finer sediments, washed out into the basin, became incorporated in detritus from other sources, to form the deposits of the valley flat (McLaughlin, 1946).

The Graters and Perkasie members persist right into the area of interfingering Brunswick shale and fanglomerate, where they merge with the transitional sandstone. (*McLaughlin*)

77. *Column 90. Somerset and Middlesex counties, New Jersey.*—The Stockton Formation is overlapped, and in most places is concealed by coastal plain deposits. The Lockatong thins rapidly eastward, until it also passes under the overlapping Cretaceous. The Brunswick Formation is developed in great thickness and consists chiefly of red shale, except in areas near the north border, where some sandstone is interbedded. Limestone fanglomerate interfingers with beds high in the Brunswick and lies unconformably upon Paleozoic limestone and shale. The First and Second Watchung Lava Flows are near the top of the Brunswick (Kummel, 1898; Darton and Kummel, 1909). (*McLaughlin*)

78. *Column 91. Bergen and Passaic counties, New Jersey.*—(Kummel, 1898, Darton and Kummel, 1908.) The Stockton Formation emerges from beneath the coastal-plain cover, but its contact with the older rocks is concealed by the Hudson River. A local black shale at North Bergen has yielded many fish remains (Schaeffer, 1941). It is possibly a thin remnant of the Lockatong Formation. Aside from this occurrence, no beds identifiable with the Lockatong have been found in the area. That formation probably disappeared, for the most part, by interfingering with red shale of the Brunswick Formation under the Cretaceous cover.

The Brunswick grades northward from red shale to sandstone, so that the boundary between Stockton and Brunswick becomes somewhat vague. The basalt flows of First Watchung, Second Watchung, and Hook mountains are near the top of the Brunswick section. Fanglomerates occur in the sandstones between Second Watchung Mountain and Hook Mountain, as well as above Hook Mountain. Black shales near the top of the Brunswick at Boonton (Morris County) yielded numerous fossil fish many years ago (Eastman, 1905). (*McLaughlin*)

79. *Column 92. Rockland county, New York* (Kummel, 1899).—The strata exposed are all older than the Watchung lava flows. The base of the Stockton remains concealed by the Hudson River. It is shown in the chart rising to the right (north) in accord with the hypothesis of a shelving floor. The beds of the Stockton strike toward the crystalline rocks across the Tappan Zee, in a manner that suggests a shallowing of the basin in that direction.

The Brunswick is mainly sandstone, with

interbeds of conglomerate—essentially a border facies. At Stony Point, limestone fanglomerate in the Brunswick lies unconformably on Paleozoic limestone. This is an additional indication of northward shelving of the floor. Possibly Triassic sedimentation never extended more than a few miles northeastward of the present end of the area. (*McLaughlin*)

80. *Column 93. Connecticut and southern Massachusetts.*—Krynine (1950) proposed the terms New Haven Arkose, Meriden Formation, and Portland Arkose for the Newark Group of the southern Connecticut valley. Emerson's names for the lava flows in Southern Massachusetts have recently been extended into Connecticut (Rodgers *et al.*, 1956).

Associated with the Hampden Lava in Massachusetts near Holyoke are bodies of tuff and agglomerate called by Emerson (1891) the Granby Tuff.

(*Rodgers, Sanders, Waage, and Gregory*)

81. *Column 94. Massachusetts.*—Emerson (1891; 1898; 1917) introduced four stratigraphic names for lithologic types in the Massachusetts Triassic; Sugarloaf Arkose, Mount Toby Conglomerate, Longmeadow Sandstone, and Chicopee Shale. This list of names does not represent a stratigraphic sequence, however; thus the Longmeadow Sandstone occurs both above and below the Chicopee Shale and above and below the lava sheets. Willard (1951; 1952) in northern Massachusetts has restricted the term Sugarloaf to the beds below the Deerfield Lava and the term Mount Toby to the conglomeratic beds above it, and has named the sandstone beds above the lava the Turners Falls Sandstone (these were part of the Longmeadow Sandstone of Emerson, 1898, but were erroneously shown as Chicopee Shale on the map in Emerson, 1917). Willard believes that the Mount Toby Conglomerate rests unconformably on the Turners Falls Sandstone, but Emerson's descriptions and Willard's mapping near Turners Falls, Massachusetts, suggest that intertonguing of conglomerate and sandstone facies is a likely alternative explanation.

(*Rodgers, Sanders, Waage, and Gregory*)

82. *Columns 95–109. Alaska.*—The available data on the Triassic stratigraphy of Alaska have been very ably summarized by Martin (1916; 1926). Two other sources of more recent date, namely *Areal geology of Alaska* by P. S. Smith (1939) and *Geology of the Arctic slope of Alaska* by T. G. Payne and others (1952), are important additional sources of information. The data on the correlation chart have been compiled principally from these sources. A large number of areal and stratigraphic studies on northern Alaska by members of the U. S. Geological Survey are in the process of publication. The paleontology of the Triassic formations of northern Alaska has been done by Tappan (1951) on the Foraminifera and by Kummel on the megafossils (invertebrates). These data are briefly summarized here for the first time.

In addition to the four major syntheses listed above, since Martin's comprehensive summary (1926) several areal studies or studies of mining districts have included additional data on Triassic formations. The more important of these are: Buddington and Chapin (1929), Moffit (1930; 1936; 1938a; 1938b; 1941; 1943), Smith and Mertie (1930), Capps (1932; 1935; 1937; 1940), Mertie (1937; 1938), Reed and Coats (1942), Reed and Dorr (1942), Guild (1942), Wayland (1943), Pecora (1942), Miller (1946), Van Alstine and Black (1946).

(*Kummel*)

83. *Column 95. Gravina Island, Alaska.*—The Upper Triassic rocks of Gravina Island consist of massive limestone, black slate with thin beds of limestone, sandstone and some conglomerate, and massive conglomerates with lenses of limestone attaining a thickness of around 1600 feet (Martin, 1926, p. 68, 69). *Halobia* cf. *H. superba* is characteristic of the fauna and suggests a Carnian age. Other parts of this rock series have yielded corals which Martin (1926, p. 70) suggests are part of the lower Norian coral zone of J. P. Smith (1912; 1927).

J. P. Smith (1927) studied large faunas from a narrow peninsula between Three Mile Cove and Thompson Cove which he considered to be identical with the lower Noric coral zone of Shasta County, California. Besides a large number of species of corals, Smith lists *Halobia alaskana, Halobia dilata, Cassianella gravinaensis, Myophoria beringiana, Arcestes* sp. *Choristoceras* cf. *C. suttonense*, etc.

Muller (1936) has ably demonstrated that the various coral horizons of Middle and Late Triassic age in western North America are difficult to recognize on the basis of the coral species alone. The data on the Gravina Island Triassic faunas are incomplete but are listed to include both Carnian and Norian horizons on the basis of the data available. (*Kummel*)

84. *Column 96. Keku Straits, Alaska.*—The Triassic strata exposed on Hamilton Bay and

elsewhere on the shores and islands at the north end of Keku straits consists of both sedimentary and volcanic rocks. Martin (1926, p. 79) described these beds as follows:

> "The sedimentary rocks include conglomeratic limestones or calcareous conglomerate, calcareous sandstone or sandy limestone, black shale, and limestone. Most of the limestones are fine-grained gray rocks with some black slaty layers. The conglomerate and sandstone are restricted to the basal sedimentary beds. The volcanic rocks include green andesitic lava, breccia, and tuff, with local intercalated sedimentary beds. The lava predominantly shows a pillow structure and is in part amygdaloidal. Some of the breccia has a limestone matrix."

The upper volcanic series contains *Monotis subcircularis* and is thus late Norian in age. The underlying clastic and carbonate rocks contain the typical Carnian fauna of *Halobia* cf. *H. superba*, etc. Float blocks collected along the north shore of Hamilton Bay have yielded species related to *Spiriferina borealis* and *Trachyceras* (*Dawsonites*) *canadense*. These Martin (1926, p. 87) correlated with the *Dawsonites* or *Nathorstites* fauna of British Columbia, but McLearn (1947; 1953) places them as late Ladinian or early Carnian. J. P. Smith (1927, p. 12) expressed some doubt of Martin's correlation on the basis of these faunas. (*Kummel*)

85. *Columns 96, 99–104. Southern Alaska.*—The age of these pre-Carnian volcanic rocks is not known except that they overlie in places lower Permian formations and conformably underlie formations of Carnian age. Martin (1926, p. 6) has summarized the occurrence and stratigraphic relations of these volcanic rocks, and his remarks are quoted here:

> "The Upper Triassic sedimentary rocks described below are underlain in many places throughout the Pacific Mountain belt by volcanic rocks that include both lavas and tuffs and that have in some places been described as greenstones. These rocks include the Nikolai greenstone of the Chitina Valley, the basic lavas and tuffs of the upper Susitna Valley, the ellipsoidal lavas of Kenai Peninsula and Kodiak Island, some of the greenstones of the Iliamna-Clark Lake district, and the ellipsoidal lavas of Hamilton Bay, in southeastern Alaska. In all these districts they clearly underlie the marine Upper Triassic strata without recognizable unconformity. It is only in the Chitina Valley that the basal contact of these rocks with underlying beds of known age has been observed. Here they rest on Carboniferous tuffs, cherts, and slates. In the Kenai Peninsula, on Kodiak Island, and probably on the west shore of Cook Inlet they overlie the slaty rocks of unknown age that are described above. On Hamilton Bay they are probably underlain by lower Permian (?) limestones. These volcanic rocks may include also the greenstone associated with the Orca group of Prince William Sound, as well as part of the greenstones of southeastern Alaska. They may possibly be correlated with the lavas and tuffs beneath the Permian (?) limestone of White River, but it more probable that they are either Permian or early Triassic."

In some places the lower part of the Nikolai Greenstones contains interbedded limestones and limey tuff containing Permian fossils. A Permian age of at least the lower part of these greenstones seems established (P. S. Smith, 1939, p. 75). (*Kummel*)

86. *Column 97. Admiralty Island, Alaska.*—The Triassic section of Admiralty Island consists mainly of limestone with some shale. The stratigraphy of these Triassic rocks is incompletely known owing to structural complications. *Monotis subcircularis* has been identified in collections from the upper part of the section, indicating an upper Norian horizon. J. P. Smith (1927, p. 11) also lists the following Carnian species: *Halobia cordillerana, Halobia dalliana, Halobia septentrionalis, Halobia symmetrica, Halobia ornatissima, Halobia austriaca, Halobia distincta, Juvavites brockensis, Juvavites knowltoni, Tropites* cf. *T. johnsoni, Discotropites sandlingensis, Discotropites davisi, Discotropites mojsvarensis, Arcestes shastensis, Pinacoceras* cf. *P. rex, Aulacoceras* cf. *A. carlottense*, from various localities at Herring and Chapin Bays. (*Kummel*)

87. *Column 98. Juneau district, Alaska.*—The Gastineau Volcanic Group was first described by Martin (1926, p. 92), who, however, gives as reference, "Eakin, H. M., and Spencer, A. C., Geology and ore deposits of Juneau Alaska: U. S. Geol. Survey Bull.—in preparation", apparently never published. The group consists of andesitic lava at the base with local lenses of slate, overlain by calcareous slate containing Upper Triassic fossils, above which is andesitic tuff overlain by slate (Martin, 1916, p. 94). The thickness is approximately 5000 feet.

The fossils identified from the Gastineau Volcanics include *Halobia* cf. *H. superba, Trachyceras* (*Protrachyceras*) *lecontei, Atractites* cf. *A. phillippi*, plus some undetermined ammonites. The correlation of the volcanic members of this group with the Triassic is in doubt (P. S. Smith, 1939, p. 76). (*Kummel*)

88. *Column 99. Alaska Peninsula.*—The contorted chert beds resemble similar chert beds in the Kenai Peninsula, Cook Inlet, and on Kodiak Island. No fossils have been reported from the contorted chert beds, and its Late Triassic age is assigned on the basis of its

stratigraphic position. That is, it overlies greenstones of late Permian or Triassic age and underlies limestone, shales, and chert containing *Monotis subcircularis* of Norian age (Martin, 1926, p. 60–64). (*Kummel*)

89. *Column 100. Cook Inlet, Alaska.*—The original description of the Kamishak chert by Martin and Katz (1912) included part of the limestone beds which Martin later considered to be an older formation on the basis of the presence of Halobias similar to *Halobia superba.*

"The Kamishak chert consists of banded chert, calcareous shale, and thin limestone in the upper part, grading downward into more massive beds of dark chert. The total thickness is at least 1000 feet and probably exceeds 2000 feet. The upper shaly and calcareous part of the formation contains *Monotis subcircularis*, but the lower, more massive chert has thus far yielded no fossils" (Martin, 1926, p. 56).

(*Kummel*)

90. *Column 100. Cook Inlet, Alaska.*—In regard to the limestones below the Kamishak chert, Martin (1926, p. 54) writes:

"The Upper Triassic limestone of the west coast of Cook Inlet has been observed only at places of complex structure and generally in close association with the Kamishak chert. Its exact relations to that formation are somewhat doubtful, but it is believed to underlie the chert. It probably contains a species of *Halobia* which resembles the Alaskan species that is usually compared with *Halobia superba* Mojsisovics, and which may indicate that the limestone corresponds in general position with the Chitistone limestone and is of Carnic age."

(*Kummel*)

91. *Column 101. Kenai Peninsula, Alaska.*—The Triassic strata exposed around Port Graham, Kenai Peninsula, include limestone, both cherty and noncherty; massive black chert; and volcanic rocks, including tuff, tuffaceous conglomerate, and breccia. The stratigraphy of these beds is incompletely known, owing to lack of exposures and to structural complications. *Monotis subcircularis* is present in the upper part of the sequence. At a lower level a species of *Halobia* occurs, but the identity and affinities of this species are not known. If it is *Halobia superba*, it would indicate that at least part of the Kenai Peninsula sequence is Carnian in age. (*Kummel*)

92. *Column 102. Upper Susitna Valley, Alaska.*—The Triassic slate, tuff, arkose, and calcareous beds of the upper Susitna Valley contain *Monotis subcircularis.* These beds are correlative with the McCarthy Formation but very different in lithologic composition (Martin, 6,192 p. 40–42). They are more similar to the beds with *Monotis subcircularis* exposed on the Kenai Peninsula. (*Kummel*)

93. *Column 103. Nizina Valley, Alaska.*—The McCarthy Formation consists of shales and limestones with considerable amount of chert beds. The formation contains *Monotis subcircularis* and attains a thickness of about 3000 feet (Martin, 1926, p. 26–32). (*Kummel*)

94. *Column 103. Nizina Valley, Alaska.*—The Chitistone Limestone conformably overlies the Nikolai Greenstone and conformably underlies the Nizina Limestone. The Chitistone Limestone consists of massive bluish-gray limestone, and the overlying Nizina Limestone of thin-bedded limestone with shale. The amount of shale increases stratigraphically upward. The Chitistone and Nizina limestones in this sense are gradational and are separated on this basis. The Chitistone attains thicknesses of about 2000 feet, and the Nizina Limestone about 1000 feet. The Chitistone Limestone contains a typical Carnian fauna including *Halobia* cf. *H. superba, Tropites, Juvavites, Arcestes* (Martin, 1926, p. 13–18; J. P. Smith, 1927, p. 10). The Nizina Limestone has yielded no recognizable fossils; because of its lithologic similarity and apparent continuity with the Chitistone Limestone it is considered as also Carnian in age. (*Kummel*)

95. *Column 104. Kotsina and Kuskulana valleys, Alaska.*—The Kuskulana Formation was first proposed by Rohn (1900) as the Kuskulana Shale. This was modified to Kuskulana Formation by Moffit and Mertie (1923) and emended to include limestone strata thought to be equivalent to the Nizina Limestone as well as limestones in the overlying McCarthy formation. (*Kummel*)

96. *Column 105. Nation River, Alaska.*—The Triassic limestone and shales along the Yukon River near the mouth of the Nation River are 400 or more feet thick and appear to include three distinct faunal zones (Martin, 1926, p. 97; J. P. Smith, 1927, p. 12). The lowest faunal zone contains *Pecten yukonensis, Eumorphotis nationalis, Lima martini, Germanonautilus brooksi, Cladiscites martini, Nathorstites alaskanus, Trachyceras* cf. *T. lecontei, Rhynchonella blackwelderi, Spiriferina yukonensis,* and *Dielasma chapini.* J. P. Smith considered this fauna to be similar to the *Dawsonites* fauna of British Columbia. McLearn (1947; 1953) refers to this British Columbia fauna as the *Nathorstites* fauna and considers it Late Ladinian or early Carnian in age. In higher beds fossils similar to the typical Carnian

species from the Canning River area and the Chitistone limestone occur. In the upper part of the formation occur the upper Norian *Monotis subcircularis*. (*Kummel*)

97. *Column 106. Canning River, Alaska.*—In recent field investigations in the Canning River region various parties of the U. S. Geological Survey have collected Triassic ceratites in the upper member of the Sadlerochit Formation. The collections contain the following species: *Ophiceras* cf. *O. greenlandicum*, *Proptychites* cf. *P. rosenkrantzei*, *Ophiceras* (*Lytophiceras*) cf. *commune*, *Ophiceras* cf. *O. tibeticum* *Otoceras boreale*, and *Claraia stachei*. These species are identical with or very similar to the Early Triassic (Scythian) faunas described by Spath (1930; 1935) from East Greenland. (*Kummel*)

98. *Columns 106–108. Northern Alaska.*—The Shublik Formation was named by Leffingwell (1919) for Upper Triassic strata in the Canning River area. The type locality is on Shublik Island in the Canning River. The formation is now recognized all across northern Alaska from Canning River to Cape Lisburne. The formation consists of dense bituminous limestone, dark-gray chert, clay shale, oil shale, and siltstone. A possible Early Triassic age is suggested for the lowest beds, and three faunal zones are recognized that indicate, respectively, Anisian, Carnian, and Norian ages.

Shale horizons from the lowest portion of the Shublik formation suggest the possibility of Early Triassic age. D. H. Dunkle, who examined the collections from this shale reports the following:

> "Collection 53AP2 118 is the incomplete and macerated skeleton of the marine paleoniscoid fish *Boreosomus*. This individual does not show the detailed structural features upon which specific definitions within this genus have been based and for convenience it can be tentatively referred to the genotype species *B. arcticus* Stensio. Indisputable remains of *Boreosomus*, in so far as now known, are restricted to the Lower Triassic (Otoceratan and Gyronitan)."

These shale beds underlie the Middle Triassic horizons discussed below.

The Middle Triassic (Anisian) age of the lower part of the Shublik Formation in some parts of the Colville Valley region is recognized from recent collections made by various U. S. Geological Survey parties. The following species have been recognized: *Posidonia* sp., *Hungarites* sp., *Leiophyllites* sp. indet., *Proteusites* n. sp., *Pseudaplococeras* n. sp. cf. *P. parvus* (Smith), *Hollandites?* sp., *Daonella* cf. *D. dubia*, and *Daonella* cf. *D. moussoni*.

Strata of Carnian age, probably equivalent to those containing the Chitistone faunas, are indicated by the presence of *Hoplotropites* cf *H. moffiti*, *Tropites stantoni*, *Sirenites* cf. *S. hayesi*, *Halobia* cf. *H. superba*, *Halobia cordillerana*, and *Spiriferina* cf. *S. yukonensis*.

The upper part of the Shublik Formation yields *Monotis subcircularis*, which is of late Norian age. (*Kummel*)

99. *Column 109. Seward Peninsula, Alaska.*—The presence of Middle Triassic (Anisian) rocks on Brooks Mountain on Seward Peninsula is inferred from rock specimens sent E. M. Kindle (1911) which contained *Daonella* cf. *D. lommeli* and *Gymnotoceras* sp. The rock specimens were supposed to have been collected from black slate, between 1000 and 2000 feet above sea level, on the southeast slope of Brooks Mountain, in the western part of Seward Peninsula. Since subsequent work in this general area has failed to disclose any Triassic rocks, serious doubt has been cast on this record (Martin, 1926, p. 116–117; P. S. Smith, 1939, p. 40). However, the previous arguments on the lack of any other occurrence of Middle Triassic strata in Alaska are no longer valid with the established Anisian age of the lower part of the Shublik Formation, which also contains *Daonella*. (*Kummel*)

100. *Column 110. East Greenland.*—Imlay (1952, p. 985) summarized the data on this formation as follows:

> "The Cape Stewart formation occurs in eastern Jameson Land and near Cape Hope in southernmost Liverpool Land (Rosenkrantz, 1929, p. 143; Koch, 1929b, p. 251, 252; Hartz, 1896; Harris, 1926; 1931; 1932a; 1932b; and 1935) and consists of 90–200 meters of nonmarine sandstone and shale that includes locally some limestone and conglomerate. It contains two distinct plant assemblages of which the lower has been assigned to the latest Triassic and the upper to the basal Jurassic. Comparable deposits are reported from Trail Island and Geographical Society Island (Maync, 1947, p. 154)."
> (*Kummel*)

101. *Column 110. East Greenland.*—Stauber (1942) has thoroughly treated the Triassic stratigraphy of East Greenland. His volume contains a summary of all previous work plus much original data. The invertebrate faunas from the Triassic formations have been studied by Spath (1927; 1930; 1935). The vertebrate faunas have been reported on by Stensiö (1932) and Nielsen (1935; 1936). Nielsen (1949) has also described some very interesting trails from these Triassic strata.

Various formational names have been proposed for parts of the Triassic sequence along

East Greenland. A summary of these systems of classification and their correlation is given by Stauber (1942, p. 20, 21). None of the formational names is used here, but rather the general faunal names for the various parts of the sequence which are in general usage. The faunal zones in the standard column at the left of the chart up to the zone of *Proptychites rosenkrantsi* are the zonal divisions recognized by Spath (1935) from his study of the ammonite faunas.

Spath (1935) is followed here in considering the beds above the *Proptychites* fauna, the *Anodontophora* beds, as also Early Triassic in age. Stauber (1942) and others incline toward making these beds Middle and Late Triassic in age. (*Kummel*)

102. *Column 111. Pearyland, Greenland.*—The Pearyland Triassic sequence was discovered by J. Troelsen during the Danish Pearyland Expedition of 1948–9 (P. C. Winther and others, 1950). The faunas collected by Troelsen and his measured sections were described by Kummel (1953). The Triassic exposures measured by Troelsen are at Støre Kløft, Herlufsholn Strand. The section consists of about 600 meters of sandstones and some shale beds. The lower part of the sequence contains upper Scythian species of *Arctoceras*, *Orbiculoidea* cf. *O. spitzbergensis*, and *Terebratula* cf. *T. margaritowi*. Higher in the section the following species indicating an Anisian age are present:—*Pearylandites troelseni*, *Groenlandites nielseni*, *Parapopanoceras* cf. *P. tetsa*, *Beyrichites* cf. *B. affinis*, and *Leiophyllites* sp. (Kummel, 1953). (*Kummel*)

Index of Stratigraphic Names by Columns

Agua Zarca 31
Alcova 55, 56, 57, 59, 60, 61, 62, 64, 65, 66
Ankareh 40, 41, 54
Applegate 24
Augusta Mountain 21
Barranca 8
Bedford Canyon 10
Bissett 36
Border 78
Boscabel 77
Brock 16
Brunswick 86, 87, 88, 89, 90, 91, 92
Bull Run 78, 79
Bully Hill 16
Camp Springs 34
Candelaria 18
Cane Spring 21
Cape Stewart 110
Cedar, Annotation 20
Chesterfield, Annotation 64
Chicopee, Annotation 81
China Mountain 21
Chinle 17, 28, 30, 31, 32, 33, 34, 37, 38, 39, 42
Chitistone 103, 104
Chugwater 55, 56, 57, 58, 59, 60, 61, 62, 63, 64, 65, 66
Church Rock 38
Conowago 83
Correo 31
Crow Mountain 61, 62, 64, 66
Cualac 3
Cultus 27
Cumnock 75
Deadman 50, 51
Deerfield 94
Dekkas, Annotation 21
Dinosaur Canyon 28, 37
Dinwoody 48, 49, 50, 51, 52, 53, 54, 55, 59, 60, 61, 64, 65
Dixie Valley 21
Dockum 32, 33, 34, 35, 45, 46, 47
Dolores 43
Dun Glen 21
Dunham, Annotation 61
El Consuelo 3
Elizabeth Furnace 85
Excelsior 18
Favret 21
Gabbs 18
Gartra 40, 41
Gastineau 98
Gettysburg 79, 80, 82, 83, 84, 86
Glen Canyon 28
Glennon, Annotation 44
Goose Egg 56, 57, 62, 63, 66
Granby, Annotation 80
Grantsville 18
Grass Valley 21
Graters 87, 88, 89
Hampden 93
Hato 26
Heidlersburg 82, 83
Higham 49, 50, 51
Holbrook 29
Holyoke 93
Hook Mountain 91
Hosselkus 15, 16
Huayacocotla 5
Hurwal 25
Jelm 44, 57
Julian, Annotation 15
Kamishak 100
Kaweah, Annotation 17
Koipato, Annotations 25, 26

Kuskulana 104
Lanes 49, 51
Little Medicine 57, 62
Lockatong 86, 87, 88, 89, 90, 91
Longmeadow, Annotation 81
Lukachuki 28, 29, 30
Luning 18
Lykins 44
Mahogany 40, 41
Manassas 78, 79
Martin Bridge 26
Matzitzi 4
McCarthy 103
Meriden 93
Milton 14
Mixtepec 3
Modin 16
Moenave 28, 37
Moenkopi 17, 28, 29, 37, 38, 42
Monitor Butte 38
Moqui 29
Moss Back 38
Mount Toby 94
Natchez Pass 21
New Haven 93
New Oxford 79, 80, 82, 83, 84, 85
Newark 70, 71, 72, 73, 74, 75, 76, 77, 78, 79, 80, 81, 82, 83, 84, 85, 86, 87, 88, 89, 90, 91, 92, 93, 94
Nizina 103
Nugget 40, 41, 42, 49, 50, 51, 54, 55, 56, 57, 60, 61, 64
Osobb 21
Otterdale 77
Owl Rock 28, 29, 30, 38
Panther Canyon 21
Pekin 75
Perkasie 87, 88, 89
Petrified Forest 28, 29, 30, 37, 38
Pierce Canyon 33
Pine, Annotation 61
Pit 16
Poleo 31
Popo Agie 55, 59, 60, 61, 62, 64
Portneuf 49, 50, 51
Portland 93
Prida 21
Raspberry 21
Red Peak 55, 59, 60, 61, 62, 64, 65
Redonda 32
Robeson 86
Rock Point 28, 29, 30
Sadlerochit 106
Sailor Canyon 14
Salitral 31
Sanford 75
Santa Ana, Annotation 15
Santa Monica, Annotation 15
Santa Rosa 32, 33, 34
Saude, Annotation 61
Sheep Pen 45, 46
Shinarump 17, 28, 29, 30, 37, 38, 42
Shnabkaib 37
Shublik 106, 107, 108
Sinbad 38
Sloan Canyon 45, 46
Slollicum, Annotation 31
Spearfish 67, 68, 69, Annotation 61
Spitzenberg, Annotation 65
Springdale 28, 37
Stanaker 40, 41
Star Peak 20
Stockton 86, 87, 88, 89, 90, 91, 92
Sugarloaf 94
Swearinger 15
Talcott 93
Tecovas 34, 35
Tegucigalpa 1
Thaynes 40, 41, 48, 49, 50, 51, 52, 53, 54
Timothy 50, 51
Timpoweap 37
Tobin 21
Todos Santos 2
Trail 15
Troublesome Creek 57
Trujillo 35
Tuckahoe, Annotation 64
Turners Falls 94
Vinita 77
Virgin 37
Watchung 91
Wingate 28, 29, 30, 38, 39
Winnemucca 21
Wood 49, 50, 51
Woodside 40, 41, 49, 53, 54
Wupatki 29

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U. S. Geological Survey, Washington 25, D. C.
Manuscript Received by the Secretary of the Society, April 24, 1957

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1515-1530, 8 FIGS. NOVEMBER 1957

THERMAL EFFECTS OF THE OCEAN ON PERMAFROST

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ABSTRACT

In high latitudes the large difference between the mean annual temperature at the ground surface and in the unfrozen sediments beneath bodies of water can affect ground temperatures to depths of several hundred feet. The effect is of particular interest near the edge of the ocean where it depends upon the magnitude of the temperature difference between the land surface and ocean bottom, the thermal properties of the ground materials, and past changes in climate and/or shore-line configuration. Theoretical considerations suggest that, except where there are transgressing shore lines, permafrost to depths greater than about 100 feet beneath the ocean bottom is not to be expected at points farther than a few thousand feet offshore. Similar considerations indicate that geothermal installations along the Arctic coast can give information regarding post-Pleistocene shore-line changes.

The geothermal effects of bodies of water offer an explanation for the anomalously large outward earth-heat flow recently reported by A. D. Misener for Resolute Bay, Cornwallis Island, N. W. T., Canada.

CONTENTS

TEXT

	Page
Introduction and acknowledgments	1515
Outline of the problem	1516
Refinements of the problem	1521
Temperature and depth distribution of permafrost near a shore line	1522
Effect of the ocean on geothermal gradients	1524
Conclusion	1528
References cited	1529

ILLUSTRATIONS

Figure	Page
1. Conditions of the idealized problem	1516
2. The function $\Psi\left(\frac{x}{z}, m\right)$	1520
3. Lines of equal ocean disturbance	1522
4. Calculated distribution of permafrost in the vicinity of the shore line for the special case outlined in text	1523
5. Calculated rise of the bottom of permafrost after a rapid transgression	1523
6. Temperatures at the Resolute Bay thermal installation with steady-state corrections	1525
7. Calculated transient gradient anomaly due to the ocean at the Resolute Bay thermal installation	1526
8. Temperatures at the Resolute Bay thermal installation with corrections for a rapid regression about 7500 years ago	1527

TABLES

Table	Page
1. The function $\Psi\left(\frac{x}{z}, m\right)$ for calculation of the geothermal disturbance due to the ocean, $x/z \geq 0$	1518
2. The function $\Psi\left(\frac{x}{z}, m\right)$ for calculation of the geothermal disturbance due to the ocean, $x/z \leq 0$	1519

INTRODUCTION AND ACKNOWLEDGMENTS

In high latitudes large bodies of water that do not freeze to the bottom in winter can have a profound effect upon the temperature and distribution of permafrost. This effect arises from the fact that the mean annual temperature of the bottom sediments in such bodies of water must be greater than the freezing temperature of the water, and the mean annual temperature of the emergent land surface may be 10°–20°C. lower. The sites of bodies of water are thus relative warm regions of the surface, and they give rise to anomalous conditions of heat flow in the underlying ground. An attempt is made in this report to estimate the resulting anoma-

lous subsurface temperatures by means of an idealized theoretical model. The results can be applied to estimate the distribution of permafrost[1] beneath the ocean's edge, and to the correction of geothermal gradients measured in connection with heat-flow studies. The results may also be applied to a study of post-Pleistocene shore-line movements.

Steady-state corrections for the thermal effects of bodies of water in temperate and tropical regions have been applied by Birch (1954a; 1956) in his studies of heat flow. For the present purposes, however, a theory taking account of the transient thermal disturbance is needed. Methods similar to those used below have been used by Birch (1950) in his study of the transient geothermal effects of topographic relief.

The work for this paper was carried out by the U. S. Geological Survey and was supported in part by the Office of Naval Research and the Arctic Research Laboratory; the Bureau of Yards and Docks, U. S. Navy; and the U. S. Army Snow, Ice, and Permafrost Research Establishment.

Prof. Francis Birch made helpful comments on the manuscript, and Max C. Brewer of the U. S. Geological Survey was responsible for obtaining the Alaskan temperature measurements upon which much of this paper is based. B. Vaughn Marshall and Gordon W. Greene, both of the U. S. Geological Survey, competently assisted with the numerical computation and the preparation of illustrations. The author has been benefited in this work by discussions with Clyde Wahrhaftig and David M. Hopkins of the U. S. Geological Survey, and with A. L. Washburn of Dartmouth College.

Outline of the Problem

The problem considered here is that of determining the ground-temperature disturbance due to the ocean, where the shore line is supposed to have shifted during some brief interval of time in the past, and to have since remained stable. The earth's surface is considered as the plane ($z = 0$), composed of the emergent land surface and the ocean bottom (Fig. 1). Neglecting relief in this way is justified along much of the North American and Asiatic Arctic coast because of the gently sloping continental shelf. The shore line is represented by a straight line ($z = 0$, $x = 0$), the emergent land surface by $z = 0$, $x < 0$, and the ocean bottom by $z = 0$, $x > 0$.

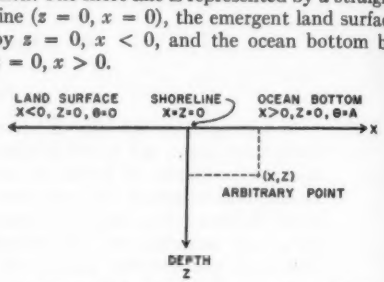

Figure 1.—Conditions of the Idealized Problem

The difference between the mean annual temperatures of the land surface and ocean bottom is represented by A, which is assumed constant for the area considered. For this to be a realistic assumption it is necessary first, that near shore at a given time the mean annual temperature of the ocean bottom and that of the land surface each be uniform, and second, that the difference between these two mean annual temperatures not change significantly with time since the shore-line movement took place. Brewer (in press) has shown that the first condition is, at present, very nearly satisfied at a typical site on the Arctic coast of Alaska. The second condition is roughly equivalent to the assumption that no large-scale systematic climatic variation has taken place since the shore-line movement. A method for removing this restriction is given in a following section.

The problem can be summarized formally as follows: find the temperature disturbance, Θ, at a subsurface point, (x, z) that satisfies the Fourier heat equation,

$$\frac{\partial^2\Theta}{\partial x^2} + \frac{\partial^2\Theta}{\partial z^2} = \frac{1}{\alpha}\frac{\partial\Theta}{\partial t}, \quad z > 0$$

subject to the following conditions:

$$\Theta(x, z, 0) = 0, \qquad t = 0, z > 0$$

$$\Theta(x, 0, t) = 0, \qquad t > 0, x < 0, z = 0$$

$$= A, \qquad t > 0, x > 0, z = 0$$

$$\Theta(x, z, t) \to 0, \qquad t > 0, x < 0, z \to \infty$$

where t = time since shore line shifted
α = thermal diffusivity of the ground materials

[1] "Permafrost" as used throughout this paper is defined as earth material whose temperature is below 0°C. for several years, regardless of its physical state. This is in keeping with the original definition proposed by Muller (1947), and is the only practical one for general studies of this kind.

The solution can be obtained by the method of Green's functions described by Carslaw and Jaeger (1947, p. 291). The Green's function for the present problem is the expression for the temperature at (x, z, t) due to an instantaneous unit line heat source at (x', z', τ) satisfying the condition that the surface $z = 0$ be maintained at zero temperature.

$$\frac{1}{4\pi\alpha(t-\tau)}\, e^{(-(x-x')^2/4\alpha(t-\tau))} \cdot \left\{e^{(-(x-x')^2/4\alpha(t-\tau))} - e^{(-(x+x')^2/4\alpha(t-\tau))}\right\}$$

This leads to

$$\Theta(x, z, t) = \frac{A}{4\pi\alpha}\int_0^t \cdot\left\{\int_0^\infty \frac{z}{(t-\tau)^2}\, e^{-(z^2+(x-x')^2/4\alpha(t-\tau))}\, dx'\right\} d\tau \tag{1}$$

Reversing the order of integration, and integrating with respect to τ gives

$$\Theta(x, z, t) = \frac{Az}{\pi}\int_0^\infty \frac{1}{z^2+(x-x')^2}\, e^{-(z^2+(x-x')^2/4\alpha t)}\, dx' \tag{2}$$

The substitution $v = \dfrac{x - x'}{z}$ reduces equation (2) to

$$\Theta(x, z, t) = \frac{A}{\pi}\int_0^\infty e^{-m(1+v^2)}\,\frac{dv}{1+v^2} + \frac{A}{\pi}\int_0^{x/z} e^{-m(1+v^2)}\,\frac{dv}{1+v^2} \tag{3}$$

where $m = \dfrac{z^2}{4\alpha t}$ [2]

[2] An alternative substitution $\tan\phi = \dfrac{x - x'}{z}$, gives the result in the form used by Birch (1950) for similar expressions:

$$\Theta(x, z, t) = \frac{A}{\pi}\int_{-(\pi/2)}^{\tan^{-1}(x/z)} e^{-(r^2/4\alpha t)}\, d\phi$$

where $r^2 = z^2 + (x - x')^2$ and $d\phi$ is the dihedral angle subtended by the strip of width dx' at the line (x, z). This form is useful for graphical approximations.

By differentiation with respect to m it can be proved that

$$\frac{1}{\pi}\int_0^\infty e^{-m(1+v^2)}\,\frac{dv}{1+v^2} = \frac{1}{\sqrt{\pi}}\int_{\sqrt{m}}^\infty e^{-u^2}\, du = \tfrac{1}{2}erfc\,\sqrt{m} \tag{4}$$

Therefore the solution sought can be written

$$\Theta(x, z, t) = A\Psi\left(\frac{x}{z}, m\right) \tag{5}$$

where

$$\Psi\left(\frac{x}{z}, m\right) = \tfrac{1}{2}erfc\,\sqrt{m} + \frac{1}{\pi}\int_0^{x/z} e^{-m(1+v^2)}\,\frac{dv}{1+v^2} \tag{6}$$

The integral in equation (6) has been tabulated by Smith (1953, p. 128) in connection with a different problem, but the range of m in his table is not suitable for most of the present applications. The function $\Psi\left(\dfrac{x}{z}, m\right)$ is presented in Tables 1 and 2 for a range of variables useful in many permafrost problems. In tabulating this function it was found that in addition to the series given by Smith, the following relation was useful owing to its rapid convergence for $m\left(\dfrac{x}{z}\right)^2 < 1$

$$\int_0^{x/z} e^{-m(1+v^2)}\,\frac{dv}{1+v^2} = \tan^{-1}\frac{x}{z} - \frac{x}{z}\left[1 - e^{-m}\right] + \frac{1}{3}\left(\frac{x}{z}\right)^3\left[1 - e^{-m}(1+m)\right] - \frac{1}{5}\left(\frac{x}{z}\right)^5\left[1 - e^{-m}\left(1 + m + \frac{m^2}{2!}\right)\right] + \cdots \approx \tan^{-1}\frac{x}{z} - m\,\frac{x}{z}, \text{ small } m$$

Thus gaps in Tables 1 and 2 for smaller values of m can be completed with the aid of

$$\Psi\left(\frac{x}{z}, m\right) \approx \tfrac{1}{2}erfc\sqrt{m} + \frac{1}{\pi}\tan^{-1}\frac{x}{z} - \frac{m}{\pi}\frac{x}{z} \approx \Psi\left(\frac{x}{z}, 0\right) - erf\sqrt{m} - \frac{m}{\pi}\frac{x}{z}$$

For larger values of m than those presented in Tables 1 and 2, $(m > 0.1)$, a table of the error

TABLE 1.—THE FUNCTION $\Psi\left(\frac{x}{z}, m\right)$ FOR CALCULATION OF THE GEOTHERMAL DISTURBANCE DUE TO THE OCEAN

$$\frac{x}{z} > 0$$

$\frac{x}{z}$ \ m	0.0000	0.0001	0.0004	0.001	0.002	0.005	0.01	0.02	0.03	0.04	0.05	0.06	0.07	0.08	0.09	0.10
0.0	0.500	0.494	0.489	0.482	0.475	0.460	0.444	0.421	0.403	0.389	0.376	0.365	0.354	0.345	0.336	0.327
0.2	0.563	0.557	0.552	0.545	0.537	0.523	0.506	0.482	0.464	0.449	0.436	0.424	0.413	0.402	0.393	0.384
0.4	0.621	0.615	0.610	0.603	0.596	0.581	0.564	0.539	0.521	0.505	0.491	0.478	0.467	0.456	0.446	0.436
0.6	0.672	0.666	0.661	0.654	0.646	0.631	0.614	0.589	0.570	0.553	0.539	0.526	0.513	0.502	0.491	0.481
0.8	0.715	0.709	0.704	0.697	0.689	0.674	0.656	0.630	0.611	0.594	0.579	0.565	0.552	0.540	0.529	0.518
1.0	0.750	0.744	0.739	0.732	0.724	0.709	0.691	0.664	0.644	0.626	0.611	0.596	0.583	0.570	0.559	0.548
1.2	0.779	0.773	0.767	0.761	0.753	0.737	0.719	0.692	0.671	0.653	0.636	0.621	0.608	0.595	0.582	0.571
1.4	0.803	0.797	0.791	0.784	0.776	0.761	0.742	0.715	0.693	0.674	0.657	0.642	0.627	0.614	0.601	0.589
1.6	0.822	0.816	0.811	0.804	0.796	0.780	0.761	0.733	0.711	0.691	0.674	0.658	0.643	0.629	0.616	0.603
1.8	0.839	0.833	0.827	0.820	0.812	0.796	0.777	0.748	0.725	0.705	0.687	0.671	0.655	0.641	0.627	0.614
2.0	0.852	0.847	0.841	0.834	0.826	0.809	0.790	0.761	0.737	0.717	0.698	0.681	0.665	0.650	0.636	0.623
2.5	0.879	0.873	0.867	0.860	0.852	0.835	0.815	0.784	0.759	0.738	0.718	0.700	0.683	0.667	0.652	0.637
3.0	0.898	0.892	0.886	0.879	0.870	0.853	0.832	0.800	0.774	0.751	0.730	0.711	0.693	0.676	0.660	0.645
4.0	0.922	0.916	0.910	0.903	0.894	0.876	0.853	0.819	0.790	0.765	0.743	0.722	0.703	0.685	0.668	0.652
5.0	0.937	0.931	0.925	0.918	0.909	0.890	0.866	0.829	0.798	0.772	0.748	0.727	0.707	0.688	0.670	0.654
6.0	0.947	0.942	0.935	0.928	0.918	0.898	0.873	0.834	0.803	0.775	0.751	0.728	0.708	0.689	0.671	0.655
8.0	0.960	0.955	0.948	0.940	0.930	0.908	0.881	0.839	0.806	0.777	0.752	0.729	0.708	0.689	0.671	0.655
10.0	0.968	0.962	0.956	0.947	0.937	0.914	0.884	0.841	0.806	0.777	0.752	0.729	0.708	0.689	0.671	0.655
∞	1.000	0.989	0.977	0.964	0.950	0.920	0.888	0.842	0.807	0.777	0.752	0.729	0.708	0.689	0.671	0.655

TABLE 2.—THE FUNCTION $\Psi\left(\frac{x}{z}, m\right)$ FOR CALCULATION OF THE GEOTHERMAL DISTURBANCE DUE TO THE OCEAN

$$\frac{x}{z} < 0$$

$\frac{x}{z}$ \ m	0.0000	0.0001	0.0004	0.001	0.002	0.005	0.01	0.02	0.03	0.04	0.05	0.06	0.07	0.08	0.09	0.10
0.0	0.500	0.494	0.489	0.482	0.475	0.460	0.444	0.421	0.403	0.389	0.376	0.365	0.354	0.345	0.336	0.327
-0.2	0.437	0.432	0.426	0.419	0.412	0.398	0.382	0.359	0.342	0.328	0.316	0.305	0.296	0.287	0.278	0.271
-0.4	0.379	0.373	0.368	0.361	0.354	0.340	0.324	0.302	0.286	0.273	0.261	0.251	0.242	0.233	0.225	0.218
-0.6	0.328	0.322	0.317	0.310	0.303	0.289	0.274	0.252	0.237	0.224	0.213	0.204	0.195	0.187	0.180	0.173
-0.8	0.285	0.279	0.274	0.267	0.260	0.246	0.231	0.211	0.196	0.184	0.173	0.164	0.156	0.149	0.142	0.137
-1.0	0.250	0.244	0.239	0.232	0.225	0.212	0.197	0.177	0.163	0.151	0.141	0.133	0.125	0.119	0.113	0.107
-1.2	0.221	0.216	0.210	0.204	0.197	0.183	0.169	0.149	0.136	0.125	0.115	0.108	0.101	0.095	0.089	0.084
-1.4	0.197	0.192	0.186	0.180	0.173	0.160	0.146	0.127	0.114	0.103	0.095	0.087	0.081	0.075	0.070	0.066
-1.6	0.178	0.172	0.167	0.160	0.154	0.140	0.127	0.109	0.096	0.086	0.078	0.071	0.065	0.060	0.056	0.052
-1.8	0.161	0.156	0.150	0.144	0.137	0.124	0.111	0.093	0.081	0.072	0.065	0.058	0.053	0.048	0.044	0.041
-2.0	0.148	0.142	0.137	0.130	0.124	0.111	0.098	0.081	0.069	0.061	0.054	0.048	0.043	0.039	0.035	0.032
-2.5	0.121	0.116	0.110	0.104	0.097	0.085	0.072	0.057	0.047	0.040	0.034	0.029	0.026	0.022	0.020	0.017
-3.0	0.102	0.097	0.092	0.086	0.079	0.067	0.056	0.042	0.033	0.027	0.022	0.018	0.015	0.013	0.011	0.009
-4.0	0.078	0.072	0.067	0.061	0.055	0.044	0.034	0.023	0.016	0.012	0.009	0.007	0.006	0.004	0.003	0.003
-5.0	0.063	0.057	0.052	0.047	0.041	0.031	0.022	0.013	0.008	0.005	0.004	0.002	0.002	0.001	0.001	0.001
-6.0	0.053	0.047	0.042	0.037	0.031	0.022	0.014	0.007	0.004	0.002	0.001	0.001	0.001	0.000	0.000	0.000
-8.0	0.040	0.034	0.029	0.024	0.019	0.012	0.006	0.002	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000
-10.0	0.032	0.026	0.022	0.017	0.013	0.007	0.003	0.001	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000
$-\infty$	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000	0.000

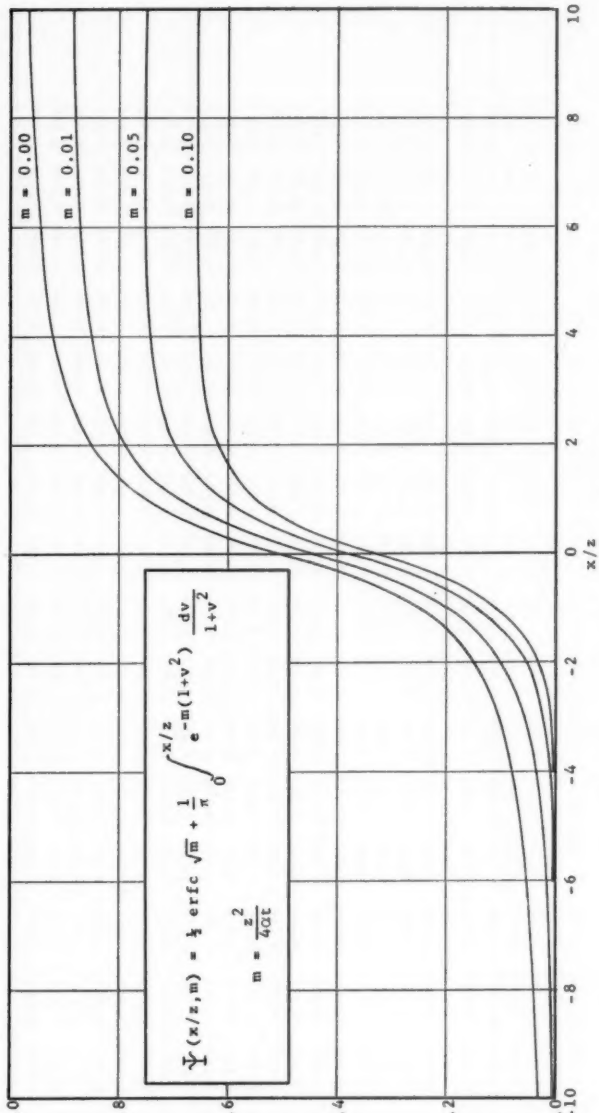

FIGURE 2.—THE FUNCTION $\Psi\left(\frac{x}{z}, m\right)$

function together with Smith's table of the integral in equation (6) can be used to calculate Ψ.

The form of the function Ψ is shown in Figure 2. The curve labelled $m = 0$ represents the steady-state result

$$\Psi\left(\frac{x}{z}, 0\right) = \tfrac{1}{2} + \tan^{-1}\frac{x}{z} \tag{7}$$

which is obtained by letting t become infinite in (6).

Refinements of the Problem

It was pointed out above that equation (5) is based on the assumption that the temperature difference, A, between land surface and ocean bottom did not change significantly since the shore-line moved. A rough estimate of A substituted in (5) should give an approximation for the temperature anomaly satisfactory for many purposes. If, however, the data warrant a more detailed analysis, the results may be easily refined. For example, if it is known, or postulated, that a climatic change at some time ($t_1 > 0$) since the shore-line movement, resulted in two relatively distinct climatic regimes, a different value of A can be used to represent the effect of each. Thus if

$$\begin{aligned} A &= A_1, & 0 < t < t_1 \\ &= A_2, & t_1 < t \end{aligned}$$

the result is

$$\Theta(x, z, t) = A_1\Psi\left(\frac{x}{z}, m\right) + (A_2 - A_1)\Psi\left(\frac{x}{z}, m_1\right), \qquad t > t_1$$

where $m_1 = \dfrac{z^2}{4\alpha(t - t_1)}$

More generally if n such climatic regimes are recognized their effects can be approximated by

$$\begin{aligned} A &= A_1, & 0 < t < t_1 \\ &= A_i, & t_{i-1} < t < t_i, \quad i = 2, 3, 4 \cdots (n-1) \\ &= A_n, & t_{n-1} < t \end{aligned}$$

The effect of the ocean is then given by

$$\Theta(x, z, t) = A_1\Psi\left(\frac{x}{z}, m\right) + \sum_{i=1}^{n-1}[A_{i+1} - A_i]\Psi\left(\frac{x}{z}, m_i\right), \qquad t > t_{n-1} \tag{8}$$

where $m_i = \dfrac{z^2}{4\alpha(t - t_i)}$

As x becomes infinite equation (8) reduces to the corresponding result for linear flow given by Birch (p. 734, equation 5, 1948).

The restriction that the shore line move suddenly, at $t = 0$, to its present position, $x = 0$, from a great distance away, can also be removed by superposition of solutions of the form of (5). Thus if three stages of shore-line movement are to be considered, the procedure is as follows: If the shore line moved from $x = \infty$ to $x = x_1$ at $t = 0$, from $x = x_1$ to $x = x_2$ at $t = t_1$, and from $x = x_2$ to $x = 0$ at $t = t_2$, the effect at a point (x, z, t) is given by

$$\begin{aligned} \Theta(x, z, t) = A\Bigg\{&\left[\Psi\left(\frac{x}{z}, m_2\right) - \Psi\left(\frac{x - x_2}{z}, m_2\right)\right] \\ &+ \left[\Psi\left(\frac{x - x_2}{z}, m_1\right) - \Psi\left(\frac{x - x_1}{z}, m_1\right)\right] \\ &+ \Psi\left(\frac{x - x_1}{z}, m\right), \qquad t > t_2 \end{aligned}$$

In the general case when n stages of shore-line movement are to be accounted for

$$\Theta(x, z, t) = A\left\{\sum_{i=1}^{n-1}\left[\Psi\left(\frac{x - x_{i+1}}{z}, m_i\right) - \Psi\left(\frac{x - x_i}{z}, m_i\right)\right] + \Psi\left(\frac{x - x_1}{z}, m\right)\right\}, \qquad t > t_{n-1} \tag{9}$$

where the shore line moved from x_i to x_{i+1} at time t_i, and $x_n = 0$. In this formula if we set $A > 0$, and take $x > 0$ on the seaward side of the present shore line, then the last term corresponds to a transgression although each term in the summation can represent either a transgression or regression.

The effect of several stages of shore-line movement combined with that of several changes in climatic regime can be obtained by combining equations (8) and (9). These equations are useful in estimating the probable uncertainty in results obtained by applying the simplified result, equation (5).

The restriction that the shore line be straight can be removed by application of results for the three-dimensional case that has been given in another paper (Lachenbruch in press). Using this method the calculation can be carried out for an arbitrary shore-line configuration, the shape of which is permitted to change with each shift in the position of the shore. Effects of concomitant climatic variation can also be accounted for by this method.

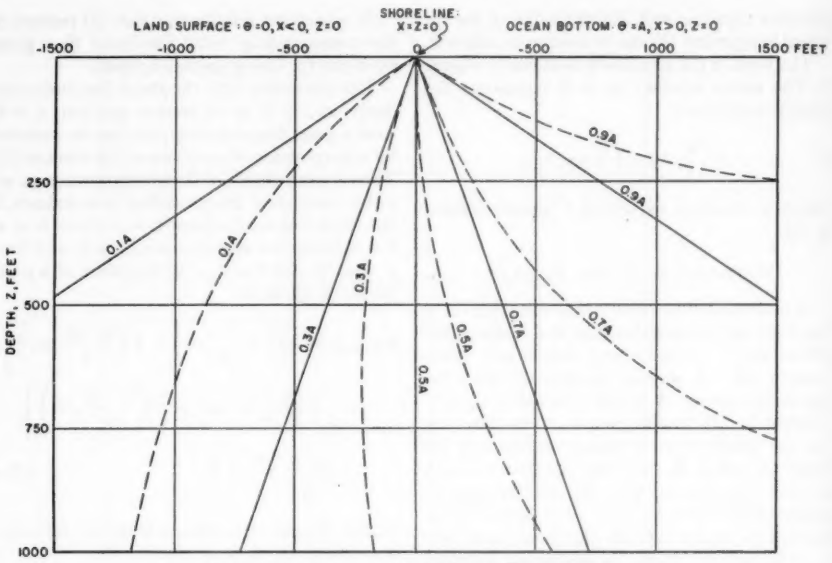

FIGURE 3.—LINES OF EQUAL OCEAN DISTURBANCE

Solid lines represent steady state; broken lines, about 10,000 years after a rapid transgression (for α = 0.01 cm²/sec).

The transient temperature disturbance beneath an infinite strip of width s whose temperature is A for $t > 0$ is given by

$$A\left\{\Psi\left(\frac{x}{z}, m\right) - \Psi\left(\frac{x-s}{z}, m\right)\right\}$$

where x is the horizontal distance from one side of the strip. This result can be used to estimate the transient thermal effect of gravel spits and rivers, or the effect of shifting the mean annual temperature by road construction.

TEMPERATURE AND DEPTH DISTRIBUTION OF PERMAFROST NEAR A SHORE LINE

If the shore line has been stationary so long that ground temperatures can be considered in equilibrium, equation (7) shows that the lines of equal temperature disturbance are the straight lines, x/z = constant, which radiate from the shoreline. These are illustrated by the solid lines in Figure 3.

To estimate how long the ocean disturbance might take to approach this equilibrium configuration we consider the case of a shore line that transgressed to its present position about 10,000 years ago ($\alpha t \approx 3 \times 10^9$ cm², assuming α = 0.01 cm²/sec). The corresponding lines of equal temperature disturbance computed from equation (5) are represented by the dashed lines in Figure 3. The disturbance is still far from equilibrium throughout most of the upper thousand feet. This suggests that the transient thermal effects of post-Pleistocene shore-line movements in high latitudes are still strongly in evidence today at relatively shallow depths.

The configuration of the 0°C isotherm, marking the bottom of permafrost, can be obtained from

$$T(z) + A\Psi\left(\frac{x}{z}, m\right) = 0 \qquad (10)$$

where $T(z)$ represents the undisturbed ground temperature characteristic of the region. As an example conditions are selected to approximate those at Barrow, Alaska, where the mean annual ocean-bottom temperature is about $-1\frac{1}{2}$°C. (Brewer, in press), the mean annual land-surface temperature is about -9°C., and the depth of undisturbed permafrost about 1300 feet (MacCarthy, 1952, p. 591). Assuming a linear variation of temperature with depth, and putting these values in (10) gives for the 0°C. isotherm

$$9\left(1 - \frac{z}{1300}\right) = 8.5\Psi\left(\frac{x}{z}, m\right), \qquad z \text{ in feet} \qquad (11)$$

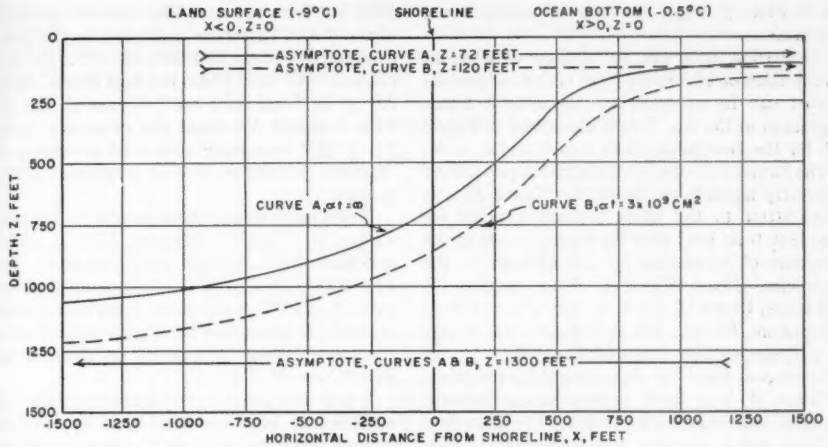
LAND SURFACE (-9°C)
X<0, Z=0
SHORELINE
OCEAN BOTTOM (-0.5°C)
X>0, Z=0
ASYMPTOTE, CURVE A, Z=72 FEET
ASYMPTOTE, CURVE B, Z=120 FEET
CURVE A, αt=∞
CURVE B, αt=3x10⁹ CM²
ASYMPTOTE, CURVES A & B, Z=1300 FEET
DEPTH, Z, FEET
HORIZONTAL DISTANCE FROM SHORELINE, X, FEET

FIGURE 4.—CALCULATED DISTRIBUTION OF PERMAFROST IN THE VICINITY OF THE SHORE LINE FOR THE SPECIAL CASE OUTLINED IN TEXT

Curve *A* represents steady state; Curve *B*, about 10,000 years after a rapid transgression (for $\alpha = 0.01$ cm²/sec).

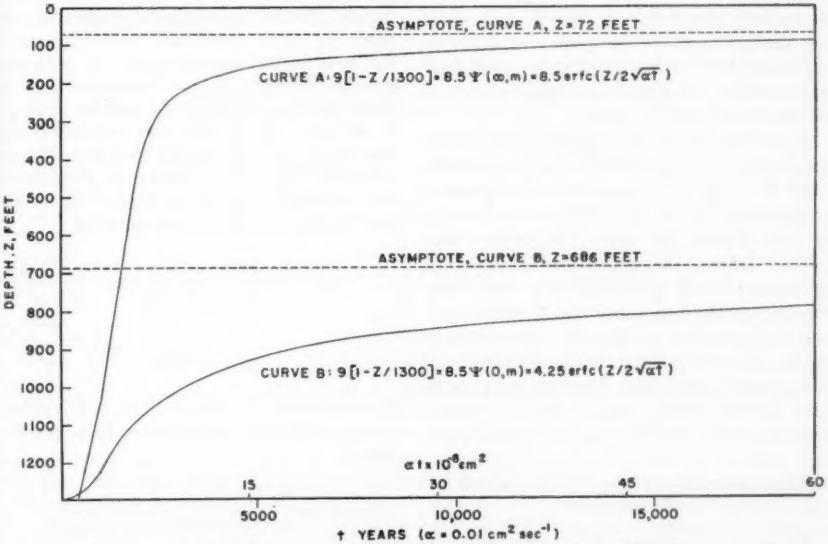
ASYMPTOTE, CURVE A, Z=72 FEET
CURVE A: 9[1-Z/1300]=8.5Ψ(∞,m)=8.5 erfc(Z/2√αt)
ASYMPTOTE, CURVE B, Z=686 FEET
CURVE B: 9[1-Z/1300]=8.5Ψ(0,m)=4.25 erfc(Z/2√αt)
DEPTH, Z, FEET
αt x 10⁹ cm²
t YEARS (α = 0.01 cm² sec⁻¹)

FIGURE 5.—CALCULATED RISE OF THE BOTTOM OF PERMAFROST AFTER A RAPID TRANSGRESSION

Curve *A*, beneath a point distant from land; Curve *B*, beneath the shore line.

The equilibrium configuration of the bottom of permafrost for these conditions is indicated by Curve *A*, Figure 4. The permafrost thins rapidly seaward and approaches a depth of about 70 feet a few thousand feet from shore. Curve *B*, Figure 4, indicates the depth distribution of permafrost to be expected about 10,000 years ($\alpha t \approx 3 \times 10^9$ cm², $\alpha = 0.01$ cm²/sec) after a rapid transgression of the sea. The equilibrium result, Curve *A*, is not affected by neglecting latent heat. The transient result, Curve *B*, however, can contain a sizable error from this sim-

plification if large amounts of moisture are present.

Equation (11) can be applied to give an indication of how rapidly the bottom of permafrost can be expected to rise after a transgression of the sea. This is illustrated in Figure 5 for the two simple cases $x = 0$ and $x = \infty$. The former corresponds to the rise of permafrost directly beneath the shore line (Curve *B*) and the latter to the effect beneath a point far enough from land that the temperatures at the bottom of permafrost are not affected by the presence of land (Curve *A*). From equation (6) it is seen that $\Psi(0, m) = \frac{1}{2}\, erfc\, \sqrt{m}$ and from equations (4) and (6) that $\Psi(\infty, m) = erfc\, \sqrt{m}$, the familiar result for linear flow. It can be shown that for the present assumptions Curve *A* is a good approximation beneath points more than 3000 feet from the shore line. Thus unless the sea has transgressed more than 3000 feet in the last few thousand years, it is unlikely that permafrost extends to depths greater than 200–300 feet at distances of a few thousand feet from shore in the Barrow area. If the sea has been stable or regressing the permafrost should thin rapidly to a depth of less than 100 feet on the order of a thousand feet from shore. These conclusions would have to be modified somewhat if appreciable amounts of latent heat were involved.

Black has presented diagrammatic cross-sections through Asia (120° E. long.) and Alaska (150° W. Long.), showing permafrost extending to depths on the order of 1000 feet at distances of tens of miles out under the Arctic Ocean (Black, 1950, p. 252). Such a distribution of permafrost would be possible only if the shore line were transgressing actively. If this is so, it can be determined by thermal observations to depths of a few hundred feet in the vicinity of the present shore line. Measurements made near Barrow, Alaska, suggest that such a distribution of permafrost is highly unlikely there.

Effect of the Ocean on Geothermal Gradients

As an example of the effect of the ocean on geothermal gradients we shall consider the heat flow study made by Misener (1955) at Resolute Bay, Cornwallis Island, N. W. T., Canada. Here the outward earth-heat flow was determined from the geothermal gradient and the thermal conductivity of core samples from the 300- to 650-foot depth interval in a well about 1200 feet from the bay. The observed gradient of about 39.5°C. per km multiplied by the mean conductivity of the dolomite cores (7.3×10^{-3} cal. $cm^{-1}\,°C^{-1}\,sec^{-1}$) lead to a heat flow of about 2.9×10^{-6} cal. $cm^{-2}\,sec^{-1}$ (Curve *A*, Fig. 6). This is almost $2\frac{1}{2}$ times the value of roughly 1.2×10^{-6} commonly considered as average or "normal" for the sea-level crust (Birch, 1954b, p. 645).

If the mean annual temperature of the land surface is $-15\frac{1}{2}$°C, (Misener, 1955, p. 1059), and that of the unfrozen sediments at the bottom of the bay is $-\frac{1}{2}$°C., a temperature difference, *A*, of 15°C. is obtained. Applying equation (5) and (7) to remove the equilibrium effect of the ocean reduces the gradient by about 30 per cent (Curve *B*, Fig. 6).

A large lake lies about 1500 feet from the well at Resolute Bay, and that part of the lake bottom that does not freeze in winter (where the water depth exceeds about 7 feet) must have a mean annual temperature greater than 0°C. Taking the mean annual temperature of the unfrozen lake bottom as +0.5°C. gives a temperature difference, *B*, of 16°C. from the land surface. We shall approximate roughly the configuration of the perennially unfrozen part of the lake bottom by the sector of a circular annulus bounded by two horizontal rays diverging from the top of the well by an angle, λ, of 0.3π, and circular arcs, concentric with the top of the well, of radii 1700 feet (R_1) and 4500 feet (R_2). The temperature disturbance, Θ_L, at a depth z in the well, at a time t after the lake was formed, is then given by

$$\Theta_L(z,t) = \frac{B\lambda}{2\pi}\left\{\frac{z}{\sqrt{z^2+R_1^2}}\, erfc\, \frac{\sqrt{z^2+R_1^2}}{2\sqrt{\alpha t}} - \frac{z}{\sqrt{z^2+R_2^2}}\, erfc\, \frac{\sqrt{z^2+R_2^2}}{2\sqrt{\alpha t}}\right\} \tag{12}$$

(Lachenbruch, in press). As $t \to \infty$ this reduces to the equilibrium result known from potential theory.

$$\Theta_L = \frac{B\lambda}{2\pi}\left\{\frac{z}{\sqrt{z^2+R_1^2}} - \frac{z}{\sqrt{z^2+R_2^2}}\right\} \tag{13}$$

Applying equation (13) to correct for the approximate equilibrium thermal effect of the lake reduces the Resolute Bay gradient to about 25.5°C. per km (Curve *C*, Fig. 6). This leads to a corrected heat flow of about 1.85×10^{-6} cal. cm^{-2} $sec.^{-1}$. Although this is a substantial reduction of the original estimate of heat flow,

the corrected value is still about 1½ times normal.

If the sea has regressed recently at Resolute Bay the equilibrium corrections applied above shall assume the sea retreated rapidly to its present position at some time, $t = 0$, following deglaciation, and that the mean annual temperature of the land surface has, in the large,

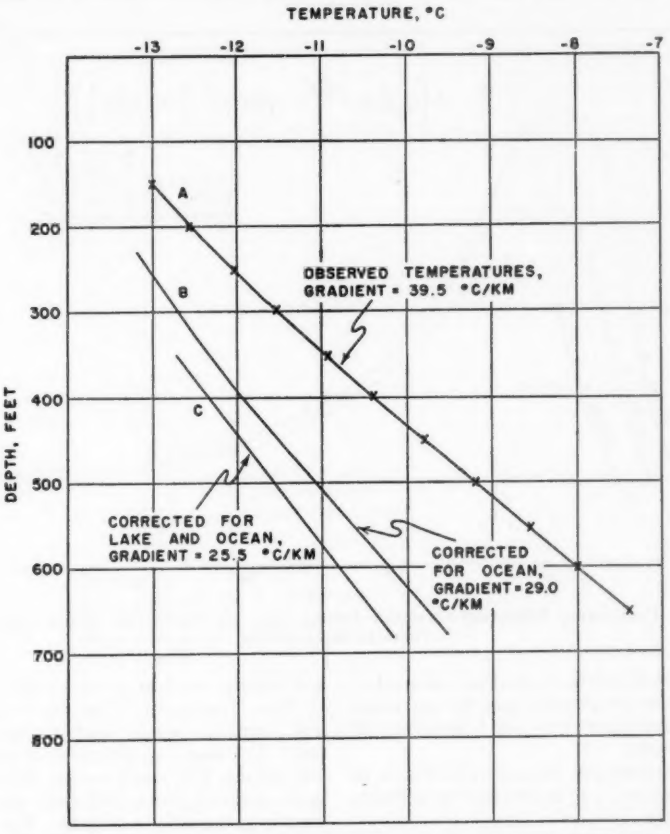

FIGURE 6.—TEMPERATURES AT THE RESOLUTE BAY THERMAL INSTALLATION WITH STEADY-STATE CORRECTIONS

are too small, and geologic evidence suggests that this is the case. Thorsteinsson and Fortier (1954, p. 17) report that strand lines occur at elevations of at least 275 feet above the present sea level at Cornwallis Island. They infer that the island had an ice cap, and that deglaciation was closely followed by uplift of the island. Thus during glaciation the land surface in the vicinity of the Resolute Bay thermal installation was probably overlain either by the sea or a great thickness of ice. In either case the mean surface temperature was probably close to 0°C. We not varied significantly from its present value since. If the position of the sea relative to the land were 125 feet higher than it is today, then the site of the Resolute Bay thermal cable would be so far offshore that the temperatures there would not be affected appreciably by the presence of the land. Thus the regression assumed is equivalent to a rapid fall in sea level of about 125 feet occurring at $t = 0$.

The mean thermal conductivity for the section under study is given by Misener as 7.3 millical per °C. cm sec. Thus the thermal

gradient associated with a normal heat flow at this locality is about 17°C. per km. From Figure 6, it is seen that the ocean correction would have to reduce the observed gradient by about 20°C. per km in order that the corrected gradient would give the normal heat-flow value. We shall now see at what time the sea would have to have receded to give a correction of this magnitude.

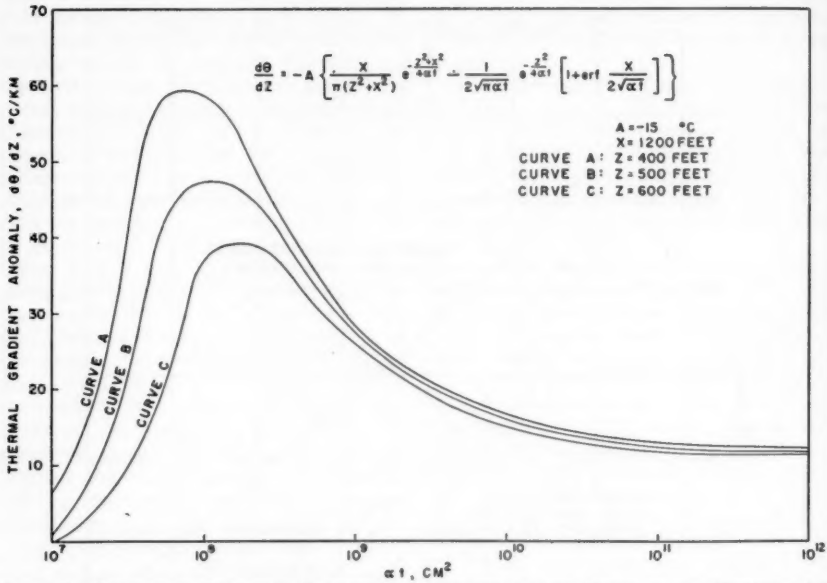

FIGURE 7.—CALCULATED TRANSIENT GRADIENT ANOMALY DUE TO THE OCEAN AT THE RESOLUTE BAY THERMAL INSTALLATION

The contribution of the ocean disturbance to the gradient at (x, z) is obtained by differentiation of (5).

$$\frac{d\theta}{dz} = -A\left\{\frac{x}{\pi(z^2+x^2)}\,e^{-(z^2+x^2)/4\alpha t} + \frac{1}{2\sqrt{\pi\alpha t}}\,e^{-z^2/4\alpha t}\left[1+\operatorname{erf}\frac{x}{2\sqrt{\alpha t}}\right]\right\} \tag{14}$$

To treat the case of a regressing sea we interchange the role of land and sea, with $x > 0$ under the land. To approximate conditions at Resolute Bay we set $A = -15$°C. and $x = 1200$ feet, the distance of the installation from the shore line. Equation (14) then gives an approximation for the gradient anomaly, due to the ocean, as a function of time for any depth, z, in the temperature well. This gradient anomaly is plotted in Figure 7 for the 400-, 500-, and 600-foot depths. It will be seen that the required gradient anomaly for all three depths occurs when αt is approximately 3×10^9 cm^2. Taking Misener's values for the conductivity and density of dolomite of 7.3×10^{-3} cal. cm^{-1} °C^{-1} sec.$^{-1}$ and about 2.75 gm cm^{-3} respectively, and assuming a specific heat of about 0.20 cal, gm^{-1} °C^{-1} leads to a diffusivity of about 0.013 cm^2 per sec. This would require that the emergence occurred about 7500 years ago, a figure that is geologically reasonable. The observed temperatures corrected for the effect of such a shore-line shift are plotted as Curve *B*, Figure 8.

The correction for the near-by lake must also be modified to fit the transient model. Assuming that the lake dates from the time of emergence, we set $\alpha t = 3 \times 10^9$ cm^2 in equation (12) to correct for the transient effect of the lake. The observed data corrected for the combined effect of lake and ocean due to a rapid emergence about 7500 years ago are plotted as Curve *C*, Figure 8. The gradient of this curve is about 17.0°C. per km which leads to a corrected earth-heat flow of about 1.25×10^{-6} cal. cm^{-2} sec^{-1}.

If this model is a reasonable approximation to conditions at Resolute Bay, more than 50 per cent of the heat flowing toward the surface there entered the earth at the ocean bottom and 5000 years ago. The two models are equally good for the present purpose in view of the uncertainty attached to the concept of normal earth-heat flow and the probable error in the

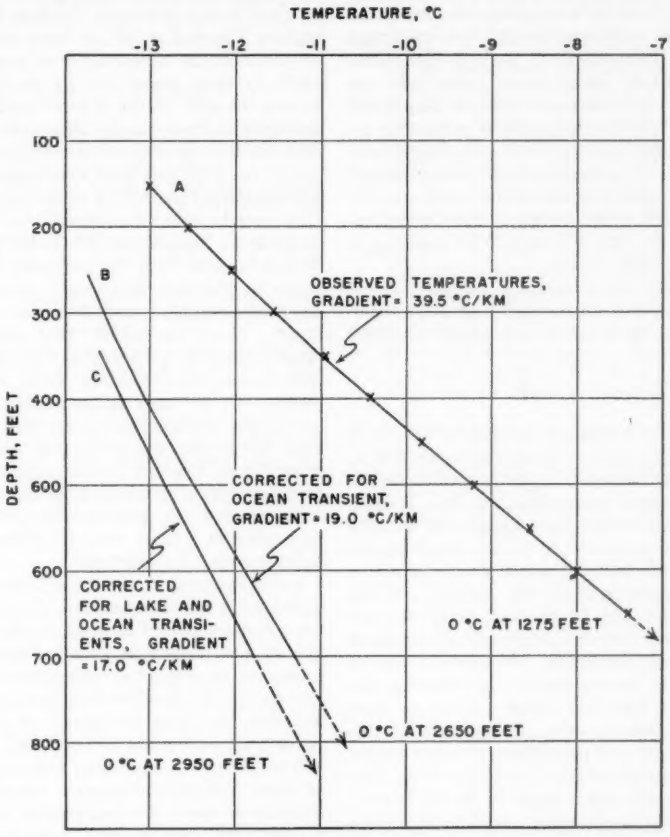

FIGURE 8.—TEMPERATURES AT THE RESOLUTE BAY THERMAL INSTALLATION WITH CORRECTIONS FOR A RAPID REGRESSION ABOUT 7500 YEARS AGO
($\alpha t = 3 \times 10^9$ cm², $\alpha = 0.013$ cm²/sec)

was not derived from sources within the earth. Similarly about 5 per cent entered through the lake bottom. Viewed another way the corrections for the ocean and lake are respectively 120 per cent and 12 per cent of steady linear earth-heat flow.

If the "normal" heat flow were considered to be about 1.1×10^{-6} cal. cm^{-2} sec.$^{-1}$ it would be associated with a "normal" gradient of about 15°C. per km which from Figures 7 and 8 is seen to correspond to a rapid emergence about mean conductivity value. If the heat flow in the area is normal, and if the sea level fell through the last 125 feet in the span of about 1000 years, these models are probably quite realistic. It is probable, however, that emergence through the last 125 feet was more gradual, and that the process is still going on. In this case the time elapsed since the sea stood at the 125-foot level is somewhat greater than that indicated by these models. In any case a mean rate of emergence through the last 125 feet on

the order of 1 or 2 feet per century is indicated. It is unlikely that the neglected effects of latent heat and climatic change would alter the order of magnitude of these results. The effects of latent heat, which can be very great in porous rocks, are probably unimportant in these dense argillaceous dolomites. Inasmuch as the actual heat flow is unknown and the post-Pleistocene history is only partly known, with only one geothermal installation at Resolute Bay it can be shown only that geologically reasonable assumptions are consistent with a local earth-heat flow within the range of normal values. A second thermal installation beneath a point interior to the island would permit a more refined estimate of the heat flow and of the sequence of post-Pleistocene events.

Differentiation of equation (14) with respect to αt reveals that the maximum gradient anomaly due to the ocean at any depth, z, occurs when

$$\alpha t = \tfrac{1}{2}z^2$$

regardless of horizontal position. This useful result is easily verified for the case of linear flow.

By linear extrapolation downward of the observed temperatures at Resolute Bay, Misener (1955) estimated the depth of permafrost at the installation to be 1280 ± 10 feet. Assuming, for the moment, that the model used above is a realistic approximation the validity of this extrapolation can be checked. Only the corrected profile (Curve *C*, Figure 8) admits linear extrapolation downward, but Curve *A* can be extended to greater depths by removing the appropriate lake and ocean corrections from the extrapolated portion of Curve *C*. By this method it is found that Misener's value for the local permafrost depth is probably in the correct range although it might be 50–100 feet too large. Under conditions of equilibrium in similar rocks beneath points interior to the island, the model suggests that the 0°C. isotherm would lie at depths of about 3000 feet (Curve *C*, Figure 8). In view of the probable glacial history of the island, however, it is highly unlikely that such equilibrium conditions are approached.

Conclusion

The theory outlined above can be easily modified to account for a gradual shore-line shift, a changing thermal regime at the surface, and topographic and other effects, but such refinements will not be warranted until more geothermal data from the arctic become available. The highly simplified model used in this paper however, is sufficient to indicate many of the general geothermal effects to be expected in the vicinity of shorelines in cold regions.

The effect of the ocean on geothermal gradients can be very great near the shore line in cold regions. This fact should be taken into account in studies of the outward flow of heat from the earth in such places. As far as the author knows, the only heat-flow value available from the Arctic is that given by Misener for Resolute Bay. Because this value is anomalously high it might be suspected that much more heat is flowing toward the earth's surface in the Arctic than elsewhere, and this would have important geodynamic implications (Birch, 1954b, p. 646). It can be seen from the foregoing discussion, however, that any geologically reasonable assumptions regarding shore-line history lead to a steady linear earth-heat flow substantially smaller than the value measured at the Resolute Bay installation. It has been shown that a particular set of assumptions consistent with geological evidence leads to an outward earth-heat flow in the range of normal values. If the heat flow at Resolute Bay is normal, the foregoing analysis suggests that the last 125 feet of emergence of this part of Cornwallis Island took place at a mean rate of 1–2 feet per century, in order of magnitude.[3]

Along the Arctic Coast of the North American mainland, a few thousand feet offshore, permafrost is not expected to occur at depths below the sea bottom greater than 2–3 hundred feet unless the shore line has undergone large transgressions in the last few thousand years. If the shoreline has been stationary, or regressing, permafrost depths are not expected to exceed 100 feet at points more than 1–2 thousand feet offshore. Valuable information regarding post-Pleistocene shore line movements can be obtained from temperature measurements to depths of a few hundred feet in the vicinity of the present shore lines. Preliminary interpretation of data from Barrow, Alaska, indicates that the thermal effect of the ocean is so great that the shore line there probably has been relatively stable for the past few thousand years. It is therefore highly unlikely that oceanic

[3] Since the initial preparation of this manuscript, a discussion of the Resolute Bay thermal anomaly has been presented by Goguel (1956, p. 183). His work is based on solutions for one-dimensional heat flow, and approximate corrections are made for the effects of latent heat. By neglecting the role of the ocean he is led to a somewhat different point of view than that expressed in this paper.

permafrost of the type indicated by Black occurs there. At Cape Simpson, however, about 60 miles east of Barrow, preliminary thermal results suggest that an active transgression might be taking place. These conclusions are consistent with geomorphic evidence from the two areas.

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U. S. Geological Survey, Menlo Park, California

Manuscript Received by the Secretary of the Society, May 3, 1957

Publication Authorized by the Director, U. S. Geological Survey

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1531–1564, 18 FIGS., 4 PLS. NOVEMBER 1957

STRUCTURE AND PETROLOGY OF THE CARIBOU LAKE INTRUSIVE BODY, ONTARIO, CANADA

BY GERALD M. FRIEDMAN

ABSTRACT

The Caribou Lake intrusive body, located in the Canadian shield, covers an area of 7 square miles. Its east-west dimension is 5½ miles; it attains a width of 2 miles but in the east tapers to a narrow constriction, 650 feet wide.

Mapping of the foliation pattern, of the topographic highs and lows which reflect the rhythmic banding in the intrusive body, and of the gneissosity in the surrounding gneisses reveals the presence of two synclinal areas in the western part of the intrusive body. The northern syncline is 2½ by 1½ miles, the southern syncline covers 1 square mile.

If the two synclinal areas represent channels through which magma ascended, then it should be possible to correlate structure with petrology. The petrological and geochemical evidence, however, points to a differentiation path that can be traced from the narrow constriction at the eastern tip of the intrusive body to the synclinal areas in the west.

This evidence is summed up:

(1) Ultrabasic rocks occur only near the eastern tip.
(2) Hypersthene decreases and augite increases as the norite is traced westward across the intrusive body.
(3) The Fe/Mg ratio in hypersthene and the Na-Si/Ca-Al ratio in plagioclase increase progressively westward.
(4) MgO decreases to the west as FeO, Fe_2O_3, and alkalies increase.
(5) Nickel, copper, and chromium decrease progressively to the west.

The intrusive body and surrounding area was mapped by aeromagnetic methods. An anomalous magnetic high of strong gradient was defined at the narrow constriction in the east where petrological and geochemical evidence indicate the presence of the funnel. This combination of observations leads to the interpretation that the channel is located at the eastern tip of the intrusive body. The significance of the two synclinal areas is a matter of conjecture. It appears that structural data alone are inadequate to determine the shapes of basic intrusive bodies.

CONTENTS

TEXT

	Page
Introduction	1532
General statement	1532
Location	1532
Acknowledgments	1532
Field study	1533
Previous work	1533
General geology	1533
Geologic setting	1533
Time of emplacement	1533
Structure	1534
Foliation pattern	1534
Linear flow	1537
Secondary lineation	1537
Roof	1537
Joints and dikes	1537
Faulting	1538
Petrography	1539
General statement	1539
Olivine-bearing rocks	1540
Pyroxenite	1540
Norite	1540
Granite pegmatites	1541
Rocks of the contact aureole bordering the pegmatites	1542
Granite and granite-gneiss	1543
Mineralization	1543
Mineralogy	1544
General statement	1544
Methods of study	1544
Plagioclase	1545
Orthopyroxene	1547
Clinopyroxene	1548
Olivine	1551
Amphibole	1551
Garnet	1552

	Page
Comparison with other intrusive bodies	1552
General statement	1552
Skaergaard intrusive body	1553
Baltimore, Maryland, gabbro	1554
Gabbro-granophyre complex, northern Wisconsin	1555
Comment	1555
Geochemistry	1556
Major elements	1556
Trace elements	1558
Geophysical surveys	1561
Airborne surveys	1561
Procedure	1561
Interpretation	1561
Ground surveys	1561
Synthesis of structural, petrological, geochemical and geophysical data	1562
References cited	1563

ILLUSTRATIONS

Figure	Page
1. Index map	1533
2. Geologic map of part of Parry Sound District, Ontario	1534
3. Geologic map showing structures in area of Caribou Lake and Memesagamesing Lake intrusive bodies	1535
4. Geologic map of Caribou Lake intrusive body showing the major structures, blocks, and compositional variation of the rock-forming minerals	1536
5. Plagioclase-orthopyroxene relationship	1546
6. Orthopyroxene-clinopyroxene relationship	1548
7. Relationship of γ' index to 2V in clinopyroxene	1549
8. Relative volumetric distribution of orthopyroxene and clinopyroxene	1550
9. Relationship of γ' index to 2V in amphibole	1552
10. Amphibole–plagioclase relationship	1552
11. Plagioclase–olivine relationship in Skaergaard intrusive body	1553
12. Plagioclase–clinopyroxene relationship in Skaergaard intrusive body	1554
13. Olivine–orthopyroxene relationship in Baltimore and Wisconsin intrusive bodies	1554
14. Orthopyroxene–clinopyroxene relationship in Baltimore gabbro and Caribou Lake intrusive rock	1555
15. Variation diagram of Caribou Lake intrusive rock	1557
16. Differentiation trend of Caribou Lake intrusive body	1558
17. Relationship of magnesium and nickel in Caribou Lake intrusive body	1560
18. Cross section showing proposed subsurface structure of Caribou Lake intrusive body	1562

Plate	Facing page
1. Geologic map of Caribou Lake intrusive body, Ontario, Canada	1531
2. Photographs of outcrops	1540
3. Photomicrographs	1541
4. Aeromagnetic map of Caribou Lake intrusive body and surrounding area	1564

TABLES

Table	Page
1. Selected modal analyses of noritic and olivine-bearing rocks and rocks of pegmatite contact aureole	1538
2. Plagioclase composition range	1545
3. Plagioclase composition distribution	1546
4. Analysis of microcline from pegmatite	1547
5. Orthopyroxene composition range	1547
6. Optical data on orthopyroxene	1547
7. Optical data on clinopyroxene	1550
8. Optical data on amphibole	1551
9. Comparative sequence in the rock-forming minerals of some basic intrusive complexes	1553
10. Differentiation trend of Caribou Lake intrusive body	1556
11. Chemical analyses of rocks	1559
12. Quantitative spectrographic analyses	1560
13. Qualitative spectrographic analyses	1560

Introduction

General Statement

Two recent contributions (Friedman 1953a; 1955) marked the beginning of a study of the basic rocks of the Parry Sound District of Ontario. The present investigation represents a continuation of this work. Some of the early results of this study have already been summarized (Friedman 1953b; 1954).

Location

The Caribou Lake complex is located in the Canadian shield in McConkey Township in the Parry Sound District of Ontario, about 2 miles west of the village of Loring (Fig. 1). It covers an area of approximately 7 square miles.

Acknowledgments

The author gratefully acknowledges the help of Mr. Richard H. Durrell in the interpretation of the aerial photographs. The Newkirk Mining Corporation kindly released the results of their geophysical investigations. Spectrographic analyses of the rocks were made by the Ontario Department of Mines, and complete chemical analyses of the major elements were performed by the Rock Analysis Laboratory of the University of Minnesota. Mrs. Richard H. Durrell and Drs. S. J. Shand, Harry Werner, and James Forgotson, Jr. read and criticized the manu-

script. Drs. H. H. Hess, Edwin Roedder and Hatten S. Yoder, Jr., suggested improvements and corrections of several sections in this paper. They should not, however, be held responsible for any defects as here presented. Dr. J. E. Armstrong made pertinent suggestions for revision of the manuscript. The research was aided by a grant-in-aid from the Sigma Xi-Resa research fund. The cost of two chemical analyses and of about 150 thin sections was defrayed by the Research Foundation of the University of Cincinnati.

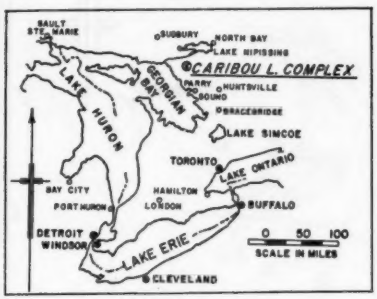

FIGURE 1.—INDEX MAP

Field Study

Two months were spent in the field mapping the intrusive body with the aid of aerial photographs. Outcrops were plotted wherever observed, but for the sake of clarity the outcrop areas were not included on the final map (Pl. 1). The dip and strike of the planes of foliation of the norites, the gneissosity of the granite gneisses, and the plunge and bearing of the lineations in the granite gneisses and norites were plotted on the map.

A reconnaissance geological map of a 100-square-mile area (Fig. 3) which includes the Caribou Lake and the near-by Memesagamesing Lake intrusive complexes was prepared from aerial photographs by Mr. J. F. Weaver. Field data collected by the author were superimposed on this map.

PREVIOUS WORK

The rocks of the Caribou Lake intrusive body have not been previously studied in detail. Coleman (1899, p. 259–262) visited the Caribou Lake area and reported the rock there to be a "...rusty diabase with more or less iron and copper pyrites and pyrrhotite". He described (1900, p. 170–171) several pits that had been sunk into the "diabase" and quoted three assays of mineralized samples. Statistical reviews of the mineral industry of Ontario (Gibson, 1909, p. 50; 1911, p. 54; 1912, p. 51; Rogers, 1916, p. 44) referred to development work on the mica properties of the Caribou Lake tract. Satterly (1943, p. 12–15) identified the basic igneous rock as a norite and recognized an ultrabasic phase of the intrusive rock near the southeast shore of Caribou Lake. He was primarily concerned with mineral deposits and described some of the abandoned feldspar claims (p. 58–59).

GENERAL GEOLOGY

Geologic Setting

Most of the country rocks into which the Caribou Lake complex has been intruded are granites and granite-gneisses. This area is part of the Grenville Province of the Canadian shield, a belt of high-grade metamorphic rocks composed for the most part of gneisses, granitic rocks, and granitized sediments.

Time of Emplacement

Oldham (1954, p. 80) presents a map prepared from aerial photographs (Fig. 2) which indicates that the observed foliation pattern of the gneisses in the area south of the Caribou Lake complex trends approximately north-south. This pattern extends for about 40–45 miles south to the city of Parry Sound. The present study establishes a similar northward-trending foliation pattern between the Wolf River fault and Woodcock Lake fault, 2–2½ miles east of Memesagamesing Lake (Fig. 3). This roughly northward-trending foliation pattern probably was brought about by the orogenic movements to which the Grenville rocks were subjected. In the area west and north of Caribou Lake this northward-trending foliation was modified by the emplacement of the complex resulting in a strike and dip pattern in the gneisses which parallels the outline of the intrusive body (Pl. 1, Figs. 3, 4), and reflects the struggle for space between the invading magma and the country rock.

Cumming *et al.* (1955) list an impressive total of 88 dated rocks from the Grenville Province. The dated samples, taken from pegmatites and veins over a large part of the Grenville Province, have an age range of 800–1100 million years. These investigators assume that the pegmatites

date periods of orogenic activity, and they assign an age of 800–1100 million years to the Grenville orogeny. The observation that the Grenville gneisses in the area under investigation have been greatly deformed and metamorphosed, whereas the Caribou Lake and near-by Memesagamesing Lake intrusive bodies show little or no deformation, suggests that the Grenville gneisses have passed through an orogenic cycle, and the basic intrusive bodies have not. Undeformed granite pegmatites which fill fractures in the Caribou Lake and Memesagamesing Lake intrusive bodies are of the same type and composition as the pegmatites dated by Cumming *et al.* (1955) and are probably of the same age. This implies that the emplacement of the intrusive bodies probably occurred between 800 and 1100 million years ago, *i.e.* in the late Precambrian. As the pegmatites and basic intrusives postdate the main Grenville orogenic activity, there is no certainty that the age of the dated granite pegmatites reflects the age of the Grenville orogeny.

The possible correlation between the Caribou Lake complex and the Sudbury norite located about 60 miles to the northwest is of economic interest. These two intrusive bodies show striking petrographic similarities. However, Cumming *et al.* (1955, p. 45) assign a somewhat greater age to the principal Sudbury ore deposits (1200 million years) and place them in the "Secondary Grenville" or "Huronian Province."

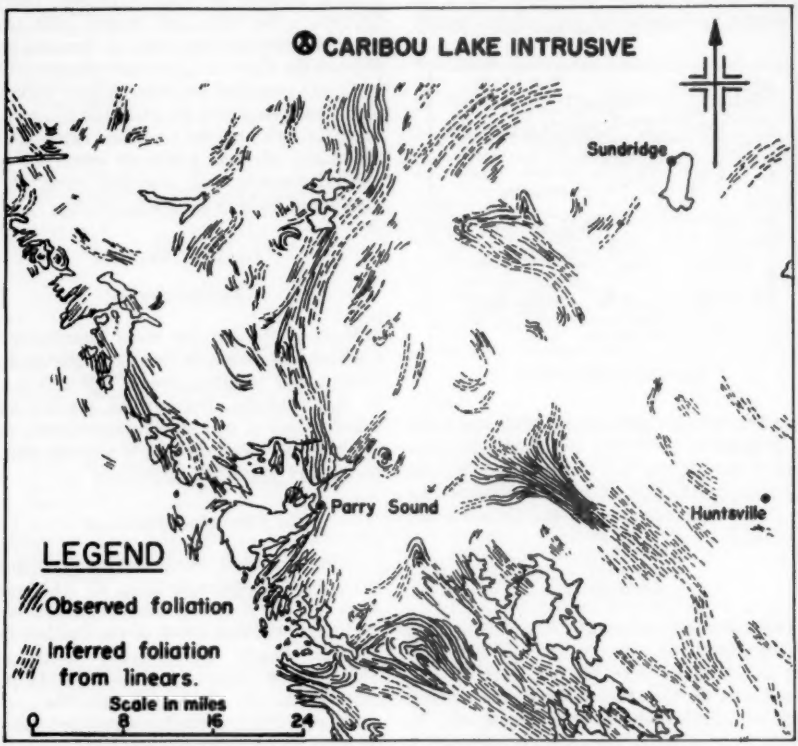

FIGURE 2.—GEOLOGIC MAP OF PART OF THE PARRY SOUND DISTRICT, ONTARIO
Shows regional foliation. Inside circle X indicates location of Caribou Lake intrusive body. After Oldham (1954, p. 80).

STRUCTURE

Foliation Pattern

Balk (1927, p. 249–303) in his classical study of the Cortlandt Complex, New York, established the relationship between the conspicuous banding in the intrusive body and its three-dimensional shape. This banding is mainly the

result of alternation of light and dark minerals and is locally so prominent that Dana (1880, p. 194–220) at first interpreted the main part of the intrusive body as a metamorphosed series of sedimentary rocks. Balk (1927) mapped the strike and dip of the planes of foliation (*i.e.* planes of banding) and suggested that the entire plutonic body is funnel-shaped, and moreover that two or possibly three smaller funnel-shaped bodies exist in the interior. The Caribou Lake norite and the Cortlandt complex show striking structural similarities.

With the aid of Balk's pioneer study it was possible to map similar structures elsewhere. Herz (1951, p. 979) concluded that the general structure of the Baltimore gabbro was that of a funnel-shaped intrusion. Friedman (1955, p. 592) developed a similar hypothesis for the Memesagamesing Lake norite mass.

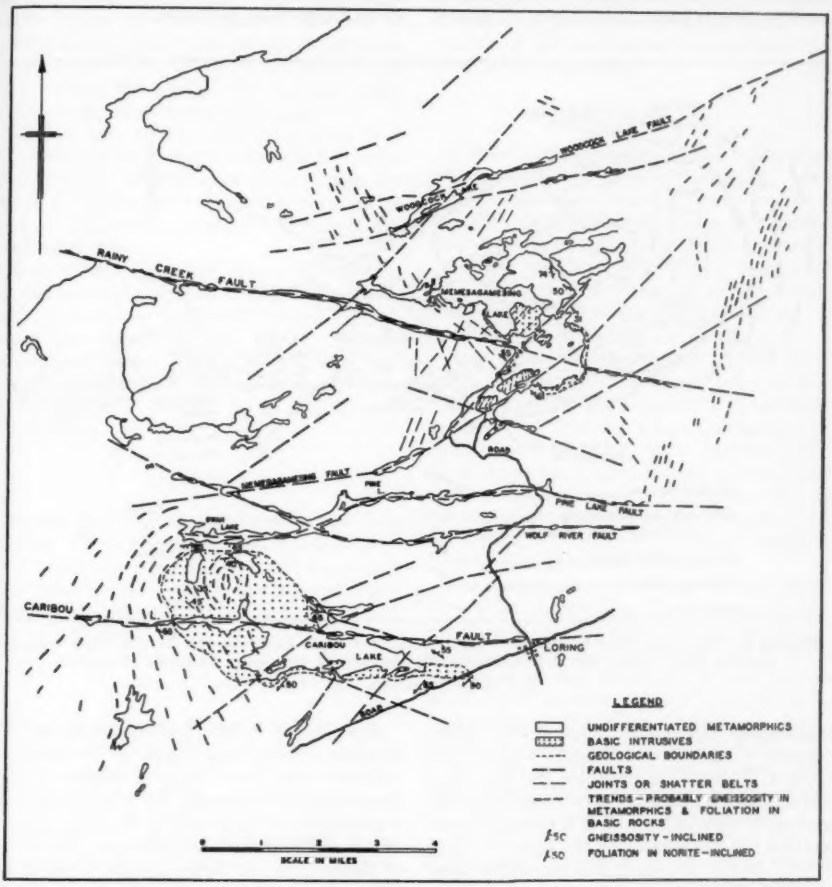

FIGURE 3.—GEOLOGIC MAP SHOWING STRUCTURES IN AREA OF CARIBOU LAKE AND MEMESAGAMESING LAKE INTRUSIVE BODIES

The alternation of dark and light bands or planes of foliation is strongly developed at Caribou Lake. In two areas the foliation dips synclinally towards focal "lows." The northern area which is roughly oval-shaped is about 2½ miles long and about 1½ miles wide (Pl. 1, fig. 4). Dips of about 30°–40° were measured on the north and west. No reliable dip readings were obtained on the south and east. A study of aerial photographs (discussed below), however,

indicates that the dips flatten on the south side close to the shores of Caribou Lake.

The second area of synclinal dips is located south and west of Peninsula Point on the south shore of Caribou Lake. It is roughly circular or oval-shaped and extends for about three-quarters of a mile in a north-south direction (Pl. 1,) fig .4. Its east-west dimension was not determined with certainty owing to insufficient dip readings at its western end. It is strongly asymmetrical; on the north side the dips are about 15°–35° and on the south side the angles range from 60°–70°. A small outcrop of gneiss is exposed on the south shore of Caribou Lake about a quarter of a mile west of Peninsula Point. This gneiss dips gently to the south (about 15°–20°) and is overlain by the basic rocks of this synclinal area (Pl. 2, fig. 1). This exposed anticlinal gneiss mass separates the northern and southern synclinal structures.

Planes of foliation are much less obvious in the eastern part of the intrusive body, although they were recognized at scattered outcrops. The gneissosity of the surrounding granite gneisses parallels the outline of the intrusive body near its western and northern part. This parallelism was observed up to distances of three-quarters of a mile from the western edge of the intrusive body. Near the eastern extremity of the complex, however, a simple structural pattern in the enclosing gneisses is absent.

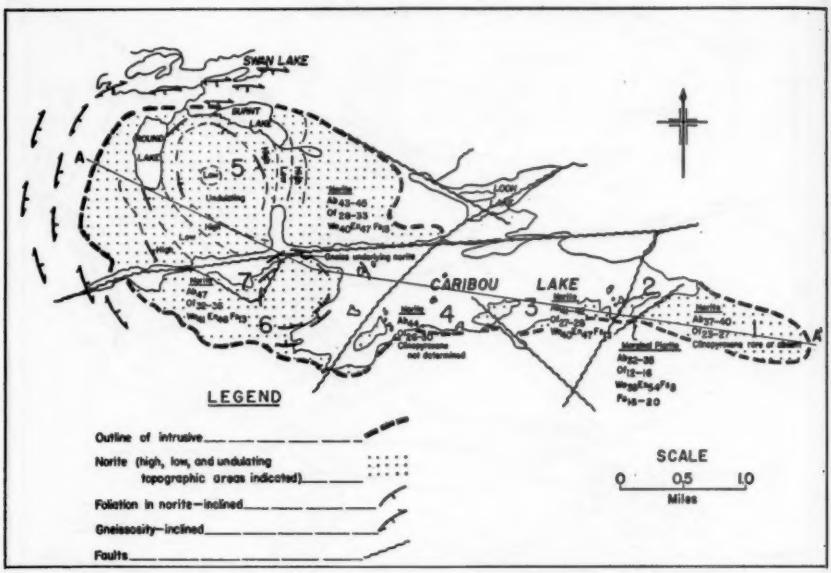

FIGURE 4.—GEOLOGIC MAP OF CARIBOU LAKE INTRUSIVE BODY SHOWING THE MAJOR STRUCTURES, BLOCKS (SEE TEXT), AND COMPOSITIONAL VARIATION OF THE ROCK-FORMING MINERALS
A–A' is line of section (Fig. 18).

A study of aerial photographs shows the presence of concentric bands in the northern synclinal area. Two bands of slightly higher topography are separated by a lower area. A topographic low is located in the approximate center of the northern synclinal area. The trend of these bands roughly parallels the strike of the planes of foliation. As the rock exhibiting this banding is all norite it seems probable that this concentric pattern of topographic highs and lows depends on the volumetric distribution of pyroxenes and feldspars. These high and low topographic areas may be the result of differential weathering and erosion. Mafic bands commonly are less resistant and usually show a greater degree of decomposition than felsic bands. The norite underlying the higher topographic areas presumably contains a lower percentage of pyroxenes than the norite underlying the lows, but this could not be confirmed in the field.

These bands which reflect the rhythmic banding so strongly developed in the intrusive body are narrow near the northern intrusive contact (about 400–500 feet in width) but widen to the south (where they are roughly 700 feet in width). This indicates a steeper dip of the bands at the northern contact and a flattening of the dips toward the south. This synclinal area therefore has an asymmetrical structure.

The southern area of synclinal dips which is located on the south shore of Caribou Lake does not show any concentric bands, although the planes of foliation are prominent in the field. Its area is considerably smaller than that of the northern syncline.

Balk (1927, p. 249–303; 1948, p. 92–95) contended that the prominent synclinal structures in basic intrusive bodies are funnels through which the magma ascended. Shand (1942, p. 409–428) and Steenland and Woollard (1952, p. 1075–1104) tested this hypothesis at Cortlandt, New York, but only partly substantiated its validity.

Linear Flow

Linear flow lines in the intrusive body are obscure. The two flow lines which were discovered plunge to the southeast and northwest. Balk (1948, p. 92) noted that evidence of linear flow is only rarely visible in funnel-shaped intrusive bodies.

Secondary Lineation

Secondary lineation is locally well developed in the metamorphic rocks into which the Caribou Lake complex has been intruded. This lineation is the result of the parallel alignment of elongate crystals, mostly feldspar, in the plane of foliation of the gneisses. The majority of the lineations plunge to the south and southeast. It has been noted that the attitude of the foliation planes near the contact with the complex is related to the emplacement of the intrusive body. It follows that the lineation trend in the same area has largely been influenced by the adjustment of the surrounding gneisses to the act of intrusion.

Roof

The roof of the intrusive body is exposed in a triangular fault block about half a mile west of Peninsula Point on the south shore of Caribou Lake. The rocks of the roof are made up of interbedded metasedimentary rocks of varying composition and include quartzites. A rock of granodioritic composition containing orthopyroxene overlies the norite and is believed to be an assimilation product between the norite magma and the metasedimentary rocks. The foliation in the rocks of the roof parallels that of the underlying norite.

Joints and Dikes

The complex is extensively jointed. The joints are apparent on aerial photographs, where measurement along strike indicates lengths of an eighth to two-thirds of a mile. The jointing is less prominently developed in the eastern extremity of the intrusive body than in the western part. Some of the joints extend out into the country rock, but the granite-gneisses are, on the whole, devoid of conspicuous joints.

Essentially two sets of joints are well developed; one roughly parallels the intrusive contact, the other has a northerly to northeasterly strike. A third set with a general westerly strike is less abundant than the other two. Joints which parallel the contact are found both in the country rock and in the complex, but are only developed near the border of the intrusive body. Many of the joints are vertical, but some have dip angles as low as 30°.

The most prominent joint set has a northerly to northeasterly strike. These joints are tensional and locally carry pyrite, pyrrhotite, graphite, and rare chalcopyrite. Owing to the absence of a linear parallelism of the minerals in the norite the term cross joints, as defined by Balk (1948, p. 27), is not applied here. Zoned granite pegmatite dikes varying in width from a few inches to 150 feet fill many of these joints. These pegmatites are restricted to the complex and were not found in the country rock.

Tension joints of the type described form at right angles to the flow of the viscous magma (Balk, 1948, p. 31). They bring about a lengthening of the igneous mass in the direction of flow. At Caribou Lake increased distention resulted from the emplacement of the numerous pegmatite dikes into the northeastward-striking joints.

The contact between the wall rock (altered norite) and the zoned granite pegmatites is sharp and straight or gently curving. Some pegmatite dikes send off smaller offshoots which follow joints that are normal to the dike. The granite-pegmatites give rise to a complex feeder system of pegmatites and granitic rocks which locally permeate the norite. Some pegmatites and granitic veins are fault-displaced. Of 49

granite pegmatite dikes that have been recorded 38 have a bearing of between N.10°E. and N.40°E. This strike direction roughly parallels that of the most prominent joint set. The pegmatites are of tabular shape, and most of them have moderately steep to very steep dips (45°–90°), but rare flat-lying pegmatites with dips as low as 20° were also found. The pegmatites show pinching and swelling relationships along their strike.

Joints in basic intrusive bodies have been studied by Balk (1927), Dally (1949, unpub. M.A. thesis, Columbia Univ.), Cserna (1950, unpub. M. A. thesis, Columbia Univ.), Van Vloten (1950, unpub. M.A. thesis, Columbia Univ.), Herz (1951), and others. Of these the more recent studies were concerned for the most part with a statistical analysis of the strike readings. It is believed that the Caribou Lake intrusive body is the first basic igneous mass in which the joints were traced systematically on aerial photographs and their exact positions determined.

Faulting

Interpretation of structure in the area studied depends considerably on geomorphic evidence. Although the presence of many faults is inferred, actual fault planes were rarely observed. The faults manifest themselves by linear depressions which locally parallel fault scarps and cut across the regional structure. They direct the courses of streams and lake shores. The only two fault planes recognized in the field are located near the southeast shore of Caribou Lake. Here the rocks of the intrusive body are offset to the south by about 100–200 feet.

Hobbs (1921, p. 507) observed that the drainage is for the most part controlled by faulting. Quirke (1936, p. 276–277), however, related the present drainage to the joint systems in the granite-gneisses. The present writer believes that some of the linear features which control the drainage are possibly tension joints of late Precambrian age, whereas others are faults. Actual displacement of the Caribou Lake intrusive body has been noted on the shores of Caribou Lake. Furthermore the linear structures, interpreted here as faults, cut across both granite-gneisses and basic intrusive rocks without regard to the joint pattern of the basic intrusives or their enclosing gneisses.

The principal faults in the area have a predominant westerly to northeasterly trend. The Caribou Lake fault has been traced for a distance of 20 miles, but is probably of much greater extent. It has an apparent strike slip of about an eighth of a mile. The Rainy Creek

TABLE 1.—SELECTED MODAL ANALYSES OF NORITIC AND OLIVINE-

Analysis No.	1	2	3	4	5	6	7	8	9	10
Rock	Picrite	Olivine Norite	Olivine Norite	Norite	Norite	Norite	Norite	Norite	Norite	Norite
Location (Block No.)	2	4	4	1	1	1	3	3	3	4
Plagioclase	7.5	33.7	40.8	62.1	47.5	63.2	69.3	71.4	45.6	56.3
Clinopyroxene	17.9	10.0	see symplektite	0.9	–	1.0	2.1	1.0	16.4	7.5
Orthopyroxene	5.9	18.3	28.8*	28.5	41.0	28.7	26.1	19.2	15.2	12.8
Olivine	11.6	13.6	0.9	–	–	–	–	–	–	–
Hornblende	–	2.0	3.9	8.0	8.3	4.8	1.2	1.8	0.4	0.2
Biotite	2.0	0.2	0.5	0.2	0.9	1.1	0.2	2.2	4.5	0.5
Serpentine	24.7	2.3	–	–	–	–	–	–	–	–
Scapolite	–	–	–	–	0.5	–	0.5	1.3	0.4	0.4
Garnet	–	7.9	–	–	1.3	1.0	0.5	2.3	15.5	13.2
Ore	3.9	1.1	0.5	0.2	0.1	0.1	–	0.1	0.8	8.0
Spinel	0.7	0.9	0.9	–	0.4	–	–	0.1	–	–
Apatite	–	–	–	–	–	–	0.1	0.1	–	0.2
Quartz	–	–	–	0.1	–	0.1	–	0.4	1.1	0.9
Carbonate	–	–	–	–	–	–	–	0.1	0.1	–
Micropegmatite	–	–	–	–	–	–	–	–	–	–
Sphene	–	–	–	–	–	–	–	–	–	–
Sericite	–	–	–	–	–	–	–	–	–	–
Zircon	–	–	–	–	–	–	–	–	–	–
Epidote	–	–	–	–	–	–	–	–	–	–
Outer Rim	19.3	–	–	–	–	–	–	–	–	–
Inner Rim	6.5	–	–	–	–	–	–	–	–	–
Symplektite	–	10.0	23.7	–	–	–	–	–	–	–
Total	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

* In this thin section it was difficult to differentiate between primary orthopyroxene and secon-

fault which cuts the Memesagamesing Lake norite mass has been traced for 16 miles. It is probably much longer. Other prominent faults include the Pine Lake fault, traced for a distance of 12 miles along strike; the Memesagamesing fault, traced for 10 miles; and the Woodcock Lake fault, traced for 10 miles. The faults appear to be normal. Several cross faults with predominant northeasterly strikes have been mapped (Fig. 3). This complex system of faults resulted in the development of a mosaic pattern of fault blocks.

The Caribou Lake intrusive body is cut by several faults. A small triangular wedge on the south shore of Caribou Lake exposes the roof. A slice of norite has been caught in a block which is now exposed on the northwest shore of Loon Lake. Slickensided norite has been discovered in the course of diamond drilling. Near slickensided surfaces the norite has been entirely serpentinized.

Bell (1890, p. 300; 1894, p. 364) was of the opinion that the weathering and erosion of diabase dikes had provided trenchlike depressions in which the present rivers flow. On the east shore of Lake Superior, in the Superior Province of the Canadian shield, the writer has seen steep-walled gorges which mark the sites of former diabase dikes. In the Caribou Lake area, however, the trench-like depressions are believed to be the sites of faults.

Friedman (1955, p. 593, 607) concludes that diabase dikes were emplaced along both Memesagamesing and Rainy Creek faults. In near-by Lount Township, Satterly (1956, p. 24) observes that some of the linear features coincide with the location of younger diabase dikes. In other Precambrian areas, such as in the Precambrian core of the Beartooth uplift of Montana and Wyoming, many of the larger linear features are filled with Precambrian diabase dikes (Spencer, 1957, p. 45).

BEARING ROCKS AND ROCKS OF PEGMATITE CONTACT AUREOLE

11	12	13	14	15	16	17	18	19	20
Norite	Norite	Norite	Norite	Norite	Norite	Norite	Amphib-olite	Amphib-olite	Anomalous contact rk.
4	5	5	5	6	6	6	2	5	5
46.0	51.4	52.9	53.8	55.8	50.4	63.9	26.0	63.5	-
22.5	13.6	20.7	12.8	12.3	20.3	16.6	-	-	-
9.8	6.1	20.4	22.4	21.1	10.5	13.5	-	-	-
-	-	-	-	-	-	-	-	-	-
5.6	7.3	5.0	8.3	8.2	11.2	2.6	65.4	31.0	38.1
1.2	5.1	0.6	2.1	-	2.7	-	2.8	3.5	39.0
-	-	-	-	-	-	-	-	-	-
-	-	-	-	1.8	-	1.6	-	-	-
7.4	11.6	-	-	-	3.9	1.3	-	-	4.8
0.7	3.9	0.3	0.5	0.6	0.5	0.2	-	0.1	0.3
-	-	-	-	-	-	-	-	-	-
-	0.1	-	-	-	0.2	-	-	1.4	9.1
1.8	0.9	0.1	0.1	0.2	0.3	0.3	5.3	-	17.7
-	-	-	-	-	-	-	-	-	-
5.0	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	0.2	-
-	-	-	-	-	-	-	-	0.3	-
-	-	-	-	-	-	-	-	trace	-
-	-	-	-	-	-	-	0.5	-	-
-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-
-	-	-	-	-	-	-	-	-	-
100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0	100.0

dary orthopyroxene which forms part of a reaction rim.

PETROGRAPHY

General Statement

The rocks which will be described are divisible into three groups:

(1) the intrusive rocks, including the ultrabasics, norites, and granite pegmatites;
(2) the rocks of the pegmatite contact aureole; and
(3) the granite-gneisses and associated metamorphic rocks into which the complex has been intruded.

Selected modal analyses of noritic and olivine-bearing rocks and of rocks of the inner contact aureole adjacent to the granite pegmatites are presented in Table 1.

Olivine-Bearing Rocks

The olivine-bearing rocks include picrite and olivine norite.

The picrite is a dark brownish-green, brown-weathering, usually coarse-grained rock in which poikilitic pryoxene plates enclose olivine crystals. Primary minerals are olivine, clinopyroxene, orthopyroxene, and plagioclase. Three-ply coronas separate olivine from feldspar. The olivine is surrounded by a rim of hypersthene, followed in order by a thin veneer of spinel, and a clinopyroxene-spinel symplektite borders the plagioclase. Precise optical data on the corona minerals are difficult to obtain. Hypersthene is aligned in prisms with the c axis normal to the periphery of the olivine, a $2V_z$ of 82° indicating a composition of about Of_{18}. The clinopyroxene has a $2V_z$ of 62° and an approximate extinction angle ($Z \wedge c$) of 41°. Plagioclase is densely clouded by minute inclusions, so that the twinning lamellae commonly cannot be recognized. There is a suggestion that these inclusions are embryo spinel crystals. Where two olivine grains are closely spaced the outer coronas merge, and plagioclase is lacking; the line of mergence is jagged and curving. Olivine crystals enclosed in pyroxene have a tendency toward roundness and give the impression that they have been resorbed during the formation of the pyroxene.

Postcorona minerals are serpentine, biotite, magnetite, and pyrite. Serpentine has replaced olivine and minerals of the inner and outer corona. The serpentine minerals are antigorite and serpophite. In places the serpentine has completely replaced the olivine, and a corona separates it from feldspar. Golden-brown spinel grains, perhaps the chrome-spinel picotite, are commonly enclosed in olivine and serpentine. Biotite has replaced pyroxene and minerals of the inner and the outer corona. Magnetite and pyrite are enclosed in biotite and are intergrown with it. Large pyroxene plates are clouded as a result of minute inclusions which produce a zoning effect. The pyroxene is deep brown where the inclusions are abundant, and it is almost colorless where they are spasmodic.

Olivine norite contains a higher feldspar content than picrite. Three-ply coronas composed of the same minerals as in picrite are also developed in the norite. Secondary hornblende and biotite enclose green intergranular spinel, an observation not made on picrite. Magnetite, pyrrhotite, pyrite, spinel, and their intergrowths are locally enclosed in secondary amphibole or mica. Four-ply coronas in which garnet forms a rim between feldspar and symplektite were also noted.

It is commonly difficult to establish whether a given pyroxene is part of the outer corona or a primary mineral. The modally analyzed picrite (Table 1) lacks spinel inclusions in the outer (clinopyroxene) rim of its coronas, and the spinel veneer between inner and outer rim is absent. In the modally analyzed olivine norite the garnet, which forms the outer rim about the olivine and which is contiguous with the plagioclase, is locally intergrown with the pyroxene layer of the outer corona and forms a symplektite with it.

Pyroxenite

The only pyroxenite occurrence noted is on the south shore of Caribou Lake adjacent to the bay opposite Caribou Island. The pyroxenite is a medium- to coarse-grained rock essentially composed of diopside and hypersthene, of which the former is dominant. Secondary minerals include garnet, amphibole, and accessory magnetite and quartz.

Norite

The norite is typically a light-gray, brown-weathering medium- to coarse-grained rock of granitoid texture. It is made up of labradorite, hypersthene, and augite (Pl. 3, fig. 1). Accessory minerals are magnetite and apatite. Norite pegmatites and aplites cut earlier norite. The pyroxene crystals in the norite pegmatite are up to 2 inches in length (Pl. 2, fig. 2). The norite has been subjected to cataclastic changes. Both feldspar and pyroxene show granulation and mortar structure and have very irregular extinction. Large crystals of these minerals are surrounded by abundant finely granulated fragments. Feldspar and pyroxene locally are bent and feldspar twinning striae obliterated. Secondary biotite also shows much bending. Cataclastic structure is more pronounced in the eastern extension of the intrusive.

PLATE 2.—PHOTOGRAPHS OF OUTCROPS

FIGURE 1.—Norite-gneiss contact. West of Peninsula Point. Kodak box 3 inches long.

FIGURE 2.—Norite-pegmatite outcrop. Island in SW. part of Caribou Lake (Blk. 6). Kodak box 3 inches long.

FIGURE 1

FIGURE 2

PHOTOGRAPHS OF OUTCROPS

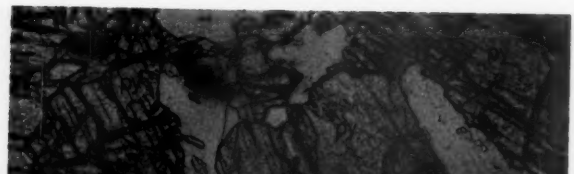

FIGURE 1

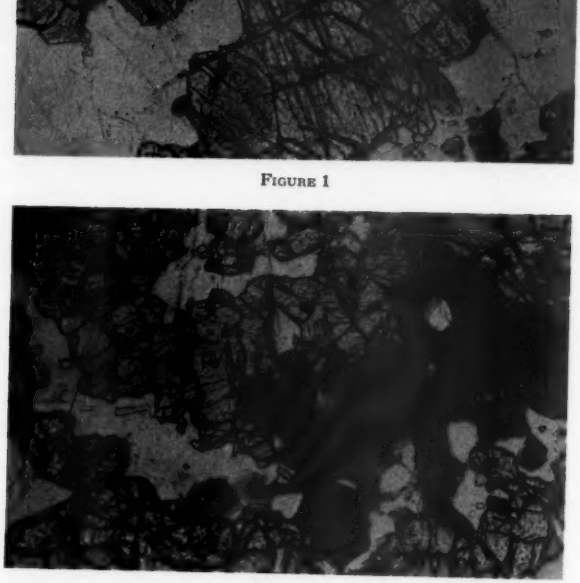

FIGURE 2

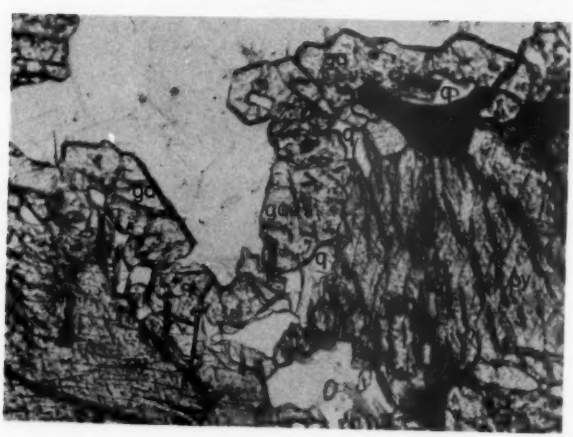

FIGURE 3

PHOTOMICROGRAPHS

Secondary minerals in the norite include hornblende, biotite, garnet, clinozoisite, scapolite, quartz, apatite, sphene, spinel, pyrite, pyrrhotite, titaniferous magnetite, chalcopyrite, and carbonate. Hornblende, biotite, and garnet are the most abundant; not a single specimen examined showed an absence of secondary minerals. Where the postmagmatic changes were least significant, hornblende, biotite, or garnet had formed at the expense of granulated pyroxene fragments (Pl. 3, fig. 2) and the larger crystals were only partially affected. In places however, even the larger pyroxenes were completely replaced. The changes were strongest close to pegmatite dikes where pyroxenes have been completely replaced by hornblende and/or biotite. The hornblende here is commonly of a blue color which is interpreted as indicating a high sodic content; the sodium has been introduced by the pegmatites.[1] All stages of gradation can be traced toward the pegmatites, from a rock with abundant pyroxene to one in which the only ferromagnesian minerals are hornblende and/or biotite.

Reaction rims of garnet commonly separate pyroxene from feldspar (Pl. 3, fig. 3). Small quartz grains are usually contiguous with the garnets or are intergrown with them. The garnet typically adjoins the feldspar, whereas the quartz is adjacent to the pyroxene. The following equation expresses the formation of this reaction rim:

$$\underset{\text{hypersthene}}{2(Mg, Fe)SiO_3} + \underset{\text{plagioclase}}{CaAl_2Si_2O_8} \rightarrow \underset{\text{garnet}}{Ca(Mg, Fe)_2\,Al_2Si_3O_{12}} + \underset{\text{quartz}}{SiO_2}$$

Disseminated opaque minerals are also bordered by garnet coronas.

In places the feldspars of the norite are clouded by minute inclusions. Many large feldspar crystals seem to have been broken up and were healed by later solutions. Vermicular quartz and plagioclase locally form a myrmekitic intergrowth. Grout (1932, p. 63) quoting Sederholm (1916) considers this type of intergrowth to be the result of deuteric action.

The opaque minerals magnetite, pyrrhotite, pyrite, and chalcopyrite are in most places enclosed in or intergrown with hornblende, biotite, or garnet, and less commonly with scapolite. Large grains of the opaque iron minerals are molded on pyroxene which they have apparently replaced. Rarely spinel is intergrown with the ore minerals. Apparently the magnetite is secondary and has formed concurrently with the late-stage silicates. Locally a sieve or diablastic texture is encountered where hornblende encloses quartz grains. This texture is, however, much more prominent in the amphibolites.

Granite Pegmatites

The granite pegmatites cutting the norite are commonly zoned. A quartz core is followed by a feldspar zone and in turn by a zone in which the feldspar is intergrown with quartz and mica. The feldspar zone, however, may be lacking. The feldspar minerals are microcline and oligoclase; one pegmatite near Burnt Lake carries moonstone. The micas include both biotite and muscovite which locally make up 2–5 per cent of the outer zone. Some of the muscovite sheets measure 1 foot in diameter.

Mica has been mined in the tract between Burnt and Caribou Lake and, it appears, that as the distance from the intrusive border of the norite increases the quality of the mica increases. The micas of the pegmatites close to Burnt Lake are too wrinkled for commercial use. On the north shore of the west arm of Caribou Lake near the big bay the muscovite is strong, free splitting, and shows least evidence of wrinkling.

In some pegmatites clear muscovite is surrounded by stained borders, caused by the inclusion of hematite or magnetite. Jahns, Griffitts, and Heinrich (1952, p. 52) term such blemishes "primary stains" as they were formed ". . . during or soon after crystallization of the mica and hence are not related to the

[1] The blue color of the hornblende does not absolutely prove a high sodium content, for a blue color is often found in ferroso-ferric compounds, as in Prussian blue and Turnbull's blue, which contain no sodium.

PLATE 3.—PHOTOMICROGRAPHS

FIGURE 1.—Norite. Pyroxene (py) and plagioclase: 82×.

FIGURE 2.—Norite. Replacement of pyroxene (py) and plagioclase by hornblende (ho), garnet (ga) and opaque minerals: 82×.

FIGURE 3.—Norite. Reaction rim of garnet (ga) and quartz (q) separating pyroxene (py) from plagioclase. Opaque mineral is magnetite: 82×.

present surface of the deposit in which they occur". Among the many types of staining and inclusions which they cite from the southeastern Piedmont (North Carolina) a clear muscovite surrounded by stained borders is not reported, on the contrary ". . . the outer parts of stained books are characteristically free from inclusions, even where the remainder of the mica is very heavily stained" (p. 55). Probably two stages of growth are involved in the formation of this mica. Formation of the early mica was followed by a period of non-growth, and the impurities were included at a later stage when growth was resumed. The contact between the stained and unstained part of the books is sharp and of typical six-sided outline.

Tourmaline is a rare accessory, but some pegmatites contain black euhedral tourmaline crystals that measure approximately 2 inches in length. Accessory garnet was noted in the zone containing the mica. Some pegmatites enclose corroded and recrystallized norite xenoliths, composed of hornblende and garnet. The axis of elongation of the xenoliths locally parallels that of the pegmatite. The pegmatite shows an alignment of minerals that may be interpreted as evidence of flow structure. This structure parallels the contact.

The adjacent amphibolite (recrystallized norite) commonly shows planes of foliation paralleling the contact of the pegmatite. Similar observations have been made by Page and others in the pegmatites of South Dakota (1953, p. 19); these investigators conclude that such features ". . . .support the suggestion that the same forces that controlled the shape of the pegmatite also deformed the adjacent wallrock".

The presence of xenoliths in the outer zone of the granite pegmatites indicates active deformation of the wall rock. Where the granite-pegmatite has succeeded in entering the metamorphic rocks of the roof of the complex the contact is undulating, and the dike widens toward the top. A prominent shear zone marks the contact between pegmatite and wall rock. Where pegmatite and recrystallized norite are contiguous the contact is sharp.

The presence of biotite and garnet and perhaps also muscovite and tourmaline in the pegmatite near its wall rock or near recrystallized norite inclusions indicates that assimilation of the norite changed the composition of the pegmatite liquid before crystallization.

A statement by Page and others (1953, p. 23) seems applicable to the Caribou pegmatites:

> "The pegmatite solution, as intruded, was a viscous liquid moving upward under pressure sufficient to deform and push aside the wallrocks and to develop gneissic and flow structures within some pegmatite units. Crystallization of the pegmatite solution was an orderly process that was interrupted at intervals by the escape of mineralizers and dissolved silicates which altered wallrocks or earlier pegmatite units. Fractional crystallization of the pegmatite solution produced zones characterized by distinctive minerals or mineral assemblages arranged in a consistent zonal sequence."

The pegmatites may be genetically related to the granite occurring on the northeast shore of Caribou Lake. It cannot be stated with certainty however, that this granite is younger than the basic complex nor that a genetic relationship exists. Ramberg (1949, p. 50) believes that ". . . most pegmatites and quartzo-feldspathic masses in folded mountain zones were formed below the solidus curve of any magma" and ". . . were formed with the help of some submagmatic metamorphic-metasomatic processes". Page and others (1953, p. 18) state that ". . . it is possible that all or part of the pegmatite solutions have been derived from metamorphic rocks by some process of ultrametamorphism". It may perhaps be significant that the mineralogical composition of the granite-pegmatites is similar to that of the granites and granite-gneisses into which the basic complex has been emplaced.

Rocks of the Contact Aureole Bordering the Pegmatites

The pegmatites are bordered by an amphibolite which is typically a feldspar-hornblende rock with varying amounts of these two minerals. In some contact aureole rocks biotite takes the place of hornblende and becomes a major constituent. Garnet is locally prominent, but as a rule it is a minor constituent or is entirely absent. In one aureole rock feldspar is lacking. Accessory minerals include garnet, clinozoisite, scapolite, sphene, magnetite, pyrite, pyrrhotite, zircon, apatite, and carbonate. This rock brings to mind the diorites and appinites of the Channel Islands described by Wells and Bishop (1955).

Under the microscope the amphibolite has a much fresher appearance than the norite, and the minerals typically are not cataclastically deformed. The feldspars usually show well-developed twin lamellae, but locally twinning is absent. Emmons and Mann (1953, p. 51) in referring to Gorai (1950, p. 149–156) report: ". . . we call attention to his conclusion that untwinned

plagioclase is found in the metasomatic type of rock". The texture is commonly diablastic (sieve texture) in which large hornblende crystals enclose small quartz grains. In the field the amphibolite commonly shows a well-developed gneissic structure.

Quartz apparently later than the feldspar locally ramifies through feldspar and cuts across the boundaries of adjacent feldspar crystals. The hornblende is mostly of the blue, sodic variety. Clinozoisite and scapolite are molded on feldspar and have formed at its expense. Quartz and feldspar locally form a myrmekitic intergrowth. The opaque minerals apparently among the last to form commonly ramify through the silicates.

The formation of the amphibolite parallels the trend of metamorphic evolution of the metadiabases of the northwest Highlands and Banff as described by Sutton and Watson (1951) and summarized by Poldervaart (1953, p. 264–265). Sutton and Watson's metamorphic sequence is considered an "abnormal" amphibolitic trend of metamorphic evolution by Poldervaart who refers to the work of Wiseman (1934) in the Scottish Highlands. The main stages in the formation of the Caribou amphibolite resemble those quoted by Poldervaart, and may be described as follows:

(1) Original texture preserved. The plagioclase is locally clouded; pyroxenes are rimmed by hornblende, biotite or garnet with intergrowth of quartz and opaque minerals; scapolite rims ores.

(2) Original texture preserved. Replacement of the primary minerals by amphiboles, micas, garnets, ores, and quartz, but relicts of original minerals still prominent; progressive zoning in the plagioclases where calcic cores are surrounded by sodic rims.

(3) Original texture modified. No relicts of primary norite minerals remain. Recrystallization of labradorite to andesine. The ferromagnesian minerals are amphibole and biotite, and less commonly garnet. The texture is diablastic with quartz inclusions in amphibole.

(4) Development of "gneissosity" or flow structure.

The widest pegmatite dike is 150 feet in width, and crops out on the south side of the west arm of Caribou Lake. It is bordered on the west by the type of amphibolite described above and in turn gives way to an amphibolite made up entirely of cummingtonite. Harker (1939, p. 134–135) considers cummingtonite to be the result of thermal metamorphism of deeply weathered basic igneous rocks exemplified by the aureole of the Land's End granite studied by Tilley and Flett (1930). At Caribou Lake the cummingtonite rock is part of the outer contact aureole of the pegmatite and has formed at the expense of the norite.

Granite and Granite-Gneiss

The country rock into which the complex intrudes is made up of red and gray granites and granite-gneisses of variable textural character. Most of them are quartzo-feldspathic rocks and contain some garnet, hornblende, and biotite. Quartz and oligoclase are the most abundant minerals; microcline is locally common. Chlorite has formed at the expense of biotite or hornblende. Magnetite, apatite, and zircon are accessory minerals. Most of the granite-gneisses are probably paragneisses. Lit-par-lit injection by pegmatite is commonplace as are discordant relationships between pegmatites and gneisses. Two textures are prominent in the gneisses, a gneissic texture, and an augen texture in which xenoblasts of feldspar are embedded in a medium- to coarse-grained matrix showing the characteristic suture texture of metamorphic rocks. Quartz xenoblasts were noted in some gneisses in which granulation appears to have taken place. A diablastic texture is developed in many gneisses and is manifested by the inclusion of quartz in feldspar xenoblasts. Locally quartz and feldspar form myrmekitic intergrowths. The cross-hatching of the microcline is barely perceptible in some of the gneisses.

Fresh-looking granite in which gneissosity is not easily seen occurs on the northeast shore of Caribou Lake.

Mineralization

Several prospect pits have been dug into the norite, notably near the contact of the intrusive with the granite-gneisses. Pyrrhotite, titaniferous magnetite, pyrite, and chalcopyrite occur either in stringers or as disseminated material in the norite. Chalcopyrite is only sparsely distributed through mineralized zones but the other three minerals are abundant.

The sulfides have formed by replacement of the norite and locally enclose corroded pyroxene relicts. Whereas the pyrite is commonly euhedral, the other minerals in the weathered material that was available for study are usually anhedral. The pyrite crystals contain inclusions of titaniferous magnetite. Chalcopyrite forms a thin intercalation between the pyrrhotite and

the magnetite. Chalcopyrite is also developed along the margin of the pyrite crystals; and magnetite is commonly intergrown with the pyrrhotite. Garnet, which is locally intergrown with quartz, borders the opaque minerals. Graphite occurs as scales in fault zones.

A chemical assay of a grab sample of weathered sulfides, analyzed by the Ontario Department of Mines, gave the following values: nickel 0.22 per cent, copper 0.32 per cent, iron 32.15 per cent, sulphur 10.08 per cent. A spectrographic analysis is presented in Table 13. A grab sample taken by Coleman (1900, p. 170–171) gave 0.37 per cent nickel; copper and platinum were not detected (analysis by the Laboratory of the School of Practical Science). Analysis of another sample gave 1.33 per cent copper, 0.32 per cent nickel, and a trace of platinum (analyst J. W. Wells). An assay of select material quoted by Coleman (1900) gave 1.55 per cent copper, 1.20 per cent nickel, 1 oz., 3 dwt. gold, and 8 g of platinum per ton.

An exploratory drilling program was carried out from the ice of Caribou Lake in the winter of 1950–1951 by Fred Chubbs and associates to determine the causative factors of anomalies previously outlined by a ground magnetometer survey. Rocks containing low values of disseminated sulfides assaying 0.226 per cent copper and 0.151 per cent nickel were encountered at a depth of 125–175 feet.

Mineralogy

General Statement

During the last few years new data have become available on the relationship between the optical properties and the chemical composition of feldspars, pyroxenes, and olivines. These data have served as a tool in the study of the compositional variation of these minerals in the differentiation of basic intrusive rocks (DeVore 1948; Walker and Poldervaart 1949; Herz 1951; Cornwall 1951; Hotz 1953; Leighton 1954; Friedman 1954). It has been known for sometime that in a fractionating basic magma feldspar becomes enriched in sodium whereas pyroxenes and olivine are enriched in iron. Buddington in 1943 (p. 123) summed up the changes:

> "In general, from near the base (of the intrusive body) upwards the plagioclases increase in soda, the olivines and hypersthenes in iron, olivine diminishes to zero, the hypersthenes diminish and augite increases, and interstitial quartz and granophyre appear in the upper facies".

Since Wager and Deer's (1939) pioneer studies the compositional changes of rock-forming minerals in the course of fractionation of basic intrusive rocks have been determined in rocks from Wyoming (DeVore 1948), South Africa (Walker and Poldervaart 1949), Maryland (Herz 1951), Pennsylvania (Hotz 1953), and Wisconsin (Leighton 1954). Hess (1941, 1949) using material from the Stillwater complex, Montana, and from other parts of the world has unravelled a major phase of the complex history of pyroxene crystallization. Yet quantitative data on the crystallization trends of the important rock-forming minerals have been fragmental and scattered. The present study is believed to be a small contribution to our knowledge of the trends of compositional variation of the major rock-forming minerals in basic plutonic complexes. The Caribou Lake intrusive body serves as an interesting example of the well-known phenomenon of enrichment in sodium in the plagioclases and of iron in the ferromagnesian minerals.

To simplify the presentation of the variation in mineral composition of the rocks of the intrusive body, reference is made throughout the ensuing discussion to the seven blocks which are shown in Figure 4. These blocks are delineated by faults and/or by the contact with the enclosing gneisses. Block 1 is located at the eastern tip of the intrusive body; blocks 5, 6, and 7 make up its western part.

Methods of Study

The refractive indices were determined by immersion methods, using white light, oils were calibrated in steps of 0.01 index, and the end point was checked on a Leitz Jelley refractometer. Estimated possible errors are ± 0.004. Optic angle determinations were made on a Leitz four-axis universal stage by direct rotation from one optic axis to the other. In this work the error is probably $\pm 0.5°$. Extinction angles were determined on the universal stage following the method of Burri (Haff 1941). Measurements are accurate to within $\pm 3°$.

The composition of the feldspars was determined on a Leitz four-axis universal stage by measuring the extinction angles perpendicular to a and correlating the values obtained with the curve of Crump and Ketner (1953, p. 31). The results were checked by determining γ' on (001), and the plagioclase composition was obtained from Crump and Ketner's curve. Determination of extinction angles with the universal stage is subject to possible errors resulting from

the rotation of the plane of polarization when the slide is tilted. Chayes (1954, p. 172–180) furthermore presents certain objections to Crump and Ketner's curves. Hess (1955, personal communication) notes that it is questionable whether (001) plates can be recognized in powders to determine γ', inasmuch as (010), (110) and $(1\bar{1}0)$ plates are also present. Troeger (1952, p. 93) on the other hand lists the (110) and $(1\bar{1}0)$ cleavages as poor.

The composition of the orthopyroxenes was determined from optic angle studies made by direct rotation from one optic axis to the other using the tables of Hess (1952, p. 180, Fig. 2). This work was supplemented by refractive index determinations. Agreement between the two methods was about ± 2 per cent.

Following the method of Walker and Poldervaart (1949, p. 631), the composition of the clinopyroxenes was determined by the intersection of the γ index curve with the curve of "normal" trend of clinopyroxene crystallization. This method of determining the composition of the clinopyroxenes was the least satisfactory optical procedure used in this study. As Walker and Poldervaart have pointed out, however, satisfactory routine optical methods for clinopyroxene determination have not been worked out; the clinopyroxene compositions cited in this paper must therefore be considered tentative until such time when better methods are available. In retrospect it should be added that Hess' (1949) clinopyroxene curves should supersede those of Walker and Poldervaart.

The olivine composition was determined from the curve of Winchell and Winchell (1951, p. 500) and checked with that of Wager and Deer (1939, p. 74).

Plagioclase

According to Hess (1955, personal communication) plagioclase feldspars in rocks of the type described always show a range in composition of ± 3 per cent Ab, and more commonly ± 5 per cent. The Ab values given in the charts (Tables 2 and 3) and used in the graph (Fig. 5) represent average feldspar values for a particular sample (thin section or powder). For the total of 41 feldspar determinations made in this study the deviations from these values averaged ± 3 per cent. These deviations may be the result of (1) experimental error and the limitations of the curves, (2) the result of the range in composition of the feldspars, and (3) a combination of these factors. In norite samples the composition range of feldspars in any one thin section usually varied from 2 to 4 per cent (14 samples), one sample showed a range of 5½ per cent, one of 9 per cent and one as high as 10 per cent. It is believed that post-magmatic chemical transformations were responsible for the high compositional variation of the last two samples. For this reason these values were not utilized in the construction of the graph (Fig. 5) which shows the plagioclase-orthopyroxene relationships.

TABLE 2.—PLAGIOCLASE COMPOSITION RANGE

Rock	Composition range	Number of samples	Remarks
Picrite	Ab_{32-35}	2	Approximation due to extreme clouding of crystals
Norite	Ab_{37-48}	13	
Norite pegmatite	Ab_{51}	1	
Norite with zoned feldspars			
core	Ab_{40-54}	6	
rim	Ab_{50-66}	6	
Norite with disseminated ore mineralization	Ab_{47-59}	4	
Rocks of pegmatite contact aureole	Ab_{47-85}	9	
Total		41	

The plagioclase crystals are generally of irregular shape, rarely lathlike. The total variation in composition is given in Table 2.

Table 3 shows the distribution of plagioclase composition in the Caribou Lake complex. The feldspars of picrite have the lowest albite content. Picrites occur only near the eastern extremity of the intrusive. Norite pegmatites contain the feldspar with the highest albite content, except for high-sulfide norites or rocks of the granite pegmatite contact aureole. Norite pegmatites were found only in the western part of the intrusive. Table 3 indicates that the feldspars in the norite show a progressive enrichment in albite westward across the intrusive body. An exception is noted in block 2 in which the feldspar is more sodic than seems consistent in view of the albite content of the feldspars in the adjoining blocks. Block 2 is a tilted fault

TABLE 3.—PLAGIOCLASE COMPOSITION DISTRIBUTION

Rock	Block number	Feldspar composition	Zoned feldspar composition	Remarks
Picrite	2	Ab_{23-35}	.....	Approximation due to extreme clouding of crystals
Norite	1	Ab_{37-40}	core Ab_{40} rim Ab_{50}	
Norite	2	Ab_{45-48}	.....	
Norite	3	Ab_{41-42}	.....	
Norite	4	Ab_{44}	.....	
Norite	5	Ab_{43-44}	core Ab_{43-51} rim Ab_{52-57}	
Norite	6	Ab_{47}	core Ab_{54} rim Ab_{66}	
Norite pegmatite	6	Ab_{51}	.....	
Amphibolite	1	Ab_{50-52}	.....	
Amphibolite	3	Ab_{47-58}	.....	
Amphibolite	4	Ab_{56-64}	.....	
Amphibolite	5	Ab_{57}	.....	
Amphibolite	6	Ab_{65}	.....	

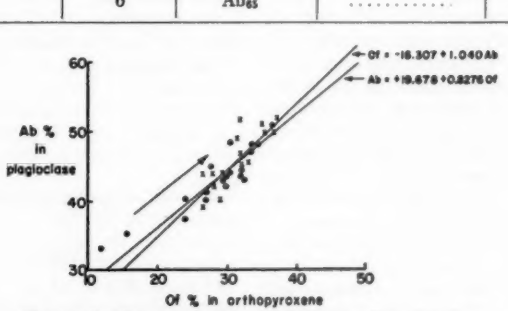

FIGURE 5.—PLAGIOCLASE-ORTHOPYROXENE RELATIONSHIP

In Caribou Lake and Memesagamesing Lake intrusive bodies. Dots represent plagioclase-orthopyroxene pairs from Caribou Lake; crosses represent pairs from the Memesagamesing Lake intrusive body which is believed to be an extension of the Caribou Lake intrusive body. The direction of progressive crystallization is indicated by arrow. Solid lines are linear regression lines.

block in which the ultrabasic rocks have been brought up and abut abruptly against the gneisses to the east and west.

It is possible to predict the plagioclase composition of a sample from its geographic location, provided the feldspar has not been changed by the action of near-by granite pegmatites. One exception to this rule, however, was noted. Norite with abundant disseminated sulfides contains a feldspar with a higher albite content than is found in non-sulfide-bearing norite. Alkali enrichment is likely to have accompanied the formation of the sulfides.

The emplacement of the granite-pegmatites caused recrystallization of the feldspar in the contact aureole. A comparison of chemical analyses of norite and amphibolite (Table 11) indicates an influx of alkalies from the pegmatites. This is also borne out by a study of the changes in the feldspar composition. The feldspars of the pegmatite aureole are invariably more sodic than the feldspars of the near-by unaltered norite.

An analysis of microcline from a granite pegmatite is given in Table 4.

Scapolite is sparsely disseminated through the norite where it has formed at the expense of the plagioclase. In the early stages of formation

it is irregular; but at a later stage it assumes a prismatic shape. It is commonly intergrown with secondary accessory sulfides or with secondary silicates, and locally forms a corona surrounding sulfide crystals. Rarely carbonate grains are associated with the scapolite.

TABLE 4.—ANALYSIS OF MICROCLINE FROM PEGMATITE
(Block 6)

Silica	65.04
Alumina	19.10
Iron (ferric)	0.13
Lime	0.25
Magnesia	0.08
Potash	13.40
Soda	1.67
Loss on ignition	0.14
Phosphorous (P_2O_5)	....
Titanium (TiO_2)	....
Total	99.81
Sp. G.	2.551
Analyst	Ontario Dept. of Mines
Reference	Satterly (1943, p. 55)

Orthopyroxene

The composition range of orthopyroxene in the different Caribou Lake rocks is given in Table 5. Table 6 shows the composition distribution and presents optical data.

The composition range of orthopyroxene in any one thin section varied from 0 to 3 per cent Of. In only a few sections, however, did this range exceed 1 per cent Of.

The orthopyroxene with the highest magnesium content is found in picrite; as the composition is traced through pyroxenite into norite the orthopyroxene becomes iron-enriched. The Fe/Mg ratio of the hypersthene in the norite increases progressively westward across the intrusive body. The most magnesian-rich hypersthene is found in block 1 and the most iron rich in block 6. As in the case of the feldspars, block 2 does not fit this pattern. It is possible to predict approximately the orthopyroxene composition of a sample from its geographic location.

There is a systematic relationship between the composition of plagioclase and orthopyroxene (Fig. 5). A plot of the orthoferrosilite content of orthopyroxene against the albite content of the plagioclase indicates a linear relationship with simultaneous iron and sodium enrichment of orthopyroxene and plagioclase respectively. Figure 5 shows the respective linear regression lines.[2] The coefficient of cor-

TABLE 5.—ORTHOPYROXENE COMPOSITION RANGE

Rock	Composition range	Number of samples	Remarks
Picrite	Of_{12-16}	2	
Pyroxenite	Of_{18}	1	
Norite	Of_{22-40}	18	
Norite with zoned hypersthene			
Core	Of_{15-21}	2	
Rim	Of_{9-15}	2	
Total	.....	25	

TABLE 6.—OPTICAL DATA ON ORTHOPYROXENE

Rock	Block	γ Index	2Vx	Estimated of	Remarks
Picrite	2		92° 85–½°	12 16	
Pyroxenite	3		81°	18	
Norite	1	1.696	71° 72° 67°	24 23 27	
	2	1.704 1.699	60–½° 67°	32 27	
	3	1.698	66-½° core 75–½ to 87° rim 85–½ to 101° 65°	27 15–22 9–15 28	Anomalous zoned orthopyroxenes
	4		68° 63°	26 30	
	5	1.703	59–½ 63–½	33 30	
	6		60° 58–½	33 34	
Norite pegmatite	6	1.709		36	

[2] A regression line of y on x is defined as the line or curve describing the relation between the mean values of y and x.

relation[3] (r) is 0.90974 which shows that the correlation is very highly significant, and it follows that about 83 per cent (coefficient of determination[4] r^2) of the variation in Of is explained by its dependence on Ab.

Clinopyroxene

The clinopyroxenes of the norite fall in the augite field, those of the ultrabasic rocks are in the endiopside field (nomenclature based on

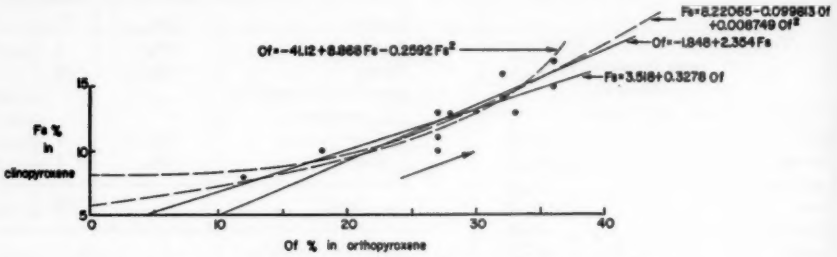

FIGURE 6.—ORTHOPYROXENE-CLINOPYROXENE RELATIONSHIP

In Caribou Lake intrusive body. The direction of progressive crystallization is indicated by arrow. Solid lines are linear regression lines, dashed lines are quadratic regression lines.

The individual points plotted in Figure 5 represent mineral pairs occurring in the same thin section. Their distribution is as follows, from lower left to upper right: the first two points represent picrite, the next olivine norite, and the remainder norite. The norite points are related in space. With few exceptions, successive points on the curve occur east of each other in the field. The points marked by X on the curve represent coexisting feldspar and orthopyroxene from the near-by Memesagamesing Lake norite mass (Friedman 1955) which is believed to be an extension of the Caribou Lake intrusive body.

Zoned orthopyroxenes are much less common than zoned feldspars and are probably magmatic in origin. They are more magnesian than unzoned orthopyroxene grains occurring in the same rock. Kuno (1936, p. 107–140) and Herz (1951, p. 987) have made similar observations in Japan and Maryland respectively.

Orthopyroxene is almost a constant constituent of the norite and is common in picrite, pyroxenite, and olivine norite. Schiller inclusions are prominent. The hypersthene is, as a rule, strongly pleochroic. The pleochroism is less pronounced in the western part of the intrusive where the orthopyroxenes are of higher iron content.

Hess, 1941, p. 516–519). The clinopyroxene is colorless in thin section. Simple twinning is commonly exhibited on (100).

Clinopyroxenes and orthopyroxenes exhibit a linear iron-enrichment relationship (Fig. 6). The correlation is highly significant (correlation coefficient, r, is 0.87684); statistically, about 77 per cent (coefficient of determination, r^2) of the variation in Of is explained by the variation in Fs. Figure 7 shows the respective regression lines. By an analysis of variance it was determined that a quadratic regression equation does not improve the fit significantly.

Table 7 gives the composition distribution of clinopyroxene in the Caribou Lake intrusive body and presents optical data.

The optic angles can be roughly correlated with the refractive indices (Fig. 7). The $2V_z$ range is from 53° in the picrites to 59° in the norite dikes. These values are above the normal optic angle range for augites which Hess (1941, p. 520) delimits as 40°–52°, with the average at about 46°. Hess (1941, p. 520) states that the

> "... optic angles of augites which have cooled slowly and thus exsolved hypersthene as fine lamellae will be higher than those in the more rapidly cooled rocks—45 to 60°. Essentially the clinopyroxene host is changed in composition from augite towards a diopside-hedenbergite pyroxene and will have comparable optical properties. The bulk composition of this clinopyroxene, plus its lamellae will, of course, be the same as that of the original augite".

The augite in the Caribou Lake complex locally carries exsolved hypersthene as fine lamellae.

[3] The coefficient of correlation is a measure of the relation between the values of two attributes of a group. The closer the coefficient is to one the greater the correlation.

[4] The coefficient of determination is a measure of the tendency of a group to assume a certain property.

An anomalous $2V_z$ value of 40.5° was measured for a clinopyroxene in a norite from the east end of block 3. The normal optic angle of the clinopyroxenes in this rock is near 56.5°.

the Stillwater complex as outlined by Hess (1940). There the earliest orthopyroxene is accompanied by a bright-green chrome-endiopside. At Caribou Lake the composition

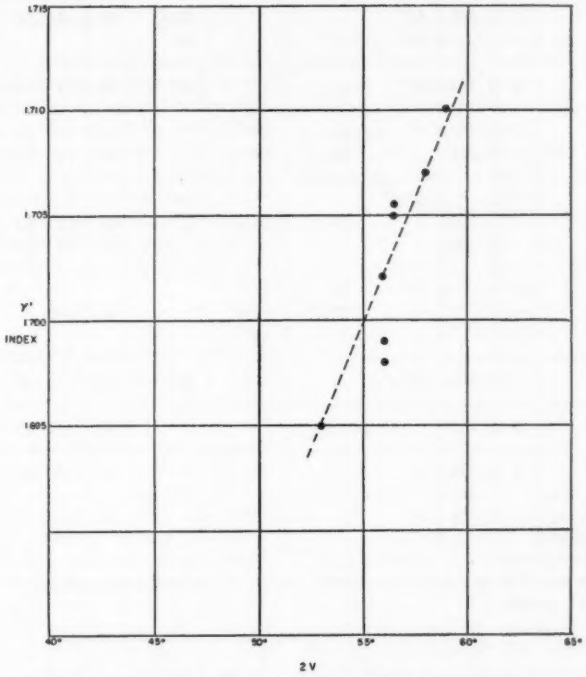

FIGURE 7.—RELATIONSHIP OF γ' INDEX TO 2 V IN CLINOPYROXENE

Zoned clinopyroxenes have cores with angles as low as 45.5°, and in one norite specimen of block 4 clinopyroxene cores have angles ranging from 50° to as low as 34°. Only one optic angle below 40° was measured. According to Hess, who sets the lower limit for augite at 32° (1941, p. 520) augites with optic angles below 40° are rare. He notes (1941, p. 527) that Walker and Poldervaart (1940) report one 31° reading on pyroxene from South Africa, but believes this value to be the "result of the presence of some minor constituent in more than normal amount" or to be "due to a small observational error". The pyroxene giving the $2V_z$ of 40.5° and the cores of the zoned pyroxenes with low optic angles are presumably relics of conditions that were maintained before the establishment of equilibrium.

The sequence of pyroxene crystallization of the Caribou Lake complex resembles that of

of the clinopyroxene of the ultrabasic rocks falls in the endiopside field. Spectrographic analyses indicate that the picrite contains 0.31 per cent chromium which is probably, at least in part, carried in the pyroxenes. The chromium content of the near-by norites ranges from a low of .009 to a high of .060 per cent chromium.

Clinopyroxene is sporadic or absent in the norite of block 1 and the east end of block 3 (as block 2 is a tilted fault block and its behavior is anomalous, it is not considered here). Farther west clinopyroxene becomes prominent (Fig. 8), and in many rocks of the northern and southern synclinal areas it becomes the second most abundant mineral (after plagioclase). At the Stillwater complex (Hess, 1940), abundant augite crystallized contemporaneously with hypersthene after the orthopyroxene reached a MgO:FeO molecular ratio of 80:20. At

TABLE 7.—OPTICAL DATA ON CLINOPYROXENE

Rock	Block	γ Index*	$2V_z$ unzoned	$2V_z$ zoned		Z∧c	Estimated			Related		
				Core	Rim		En	Fs	Wo	Of	Fa	Ab
Picrite	2	1.699	53° 55°			42½° 38°	54	8	38	12 16	16	32-35 35
Pyroxenite	3	1.702	56°			40°	52	10	38	18		
Norite	2	1.703 1.711		51° 50°	56° 58°		50 45	11 14	39 41	27 32		45 52
	3	1.709 1.710	40½°	45½°	56½°	36° 43°	47 46	13 13	40 41	15 27 28		41 42
	4		57°	34° 41½° 50°	57° 57½° 56°							
	5	1.709	56½°			45½°	47	13	40	33		44
	6	1.710					46	13	41	30		44
Norite-aplite	4	1.714	59°				42	17	41			
Norite (disseminated ore mineralization).	6	1.706	56°				49	11	40			54

* The γ-index was obtained by adding 0.004 to γ′, an approximation suggested by Walker and Poldervaart (1949, p. 630).

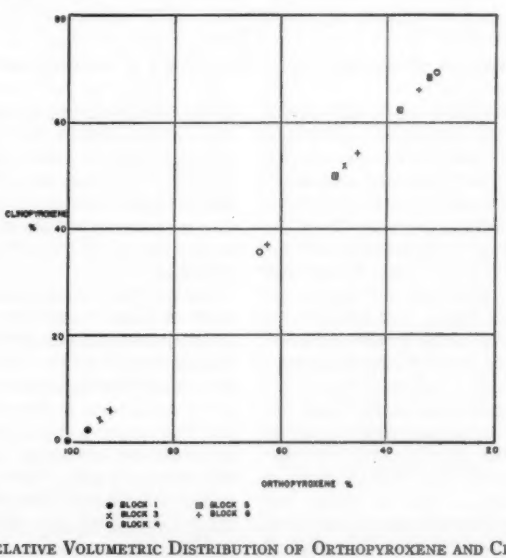

FIGURE 8.—RELATIVE VOLUMETRIC DISTRIBUTION OF ORTHOPYROXENE AND CLINOPYROXENE In norite of Caribou Lake blocks.

Caribou this ratio is about 75:25. In block 1 where clinopyroxene is sporadic or absent the range of orthopyroxene composition is about Of 21–25°.

Olivine

Olivine is not a common mineral at Caribou Lake. It is found in the picrites and in the scattered occurrences of olivine norite. The beta index of olivine in the picrite is near 1.691, $2V_z = 87°$. These figures indicate a molecular proportion of about 80 per cent forsterite in the olivine. An optic angle determination of $2V_z = 88°$ was obtained on an olivine from the olivine norite near Peninsula Point. The olivine in the picrites has been largely replaced by serpentine.

Amphibole

Amphiboles have been found in the pyroxenites, olivine norites, norites, and the rocks of the inner and outer aureole of the granite pegmatites. They are most abundant in the aureole rocks and least abundant in the pyroxenites. Amphibole was not observed in the picrites.

The amphibole varies in color and pleochroism. It is light yellow to almost colorless in pyroxenite, and yellow, yellow green, and brown in olivine norite. The color of the hornblende in the norite is mostly green, but this green is subject to variable modification and includes brownish, yellowish, and bluish greens. Locally the green color is replaced by a greenish brown.

The color of the amphiboles of the inner contact aureole is green to bluish green to blue, the amphibole of the outer aureole is a colorless cummingtonite. The amphiboles are invariably secondary. They have formed at the expense of the primary silicates of the norites and ultrabasic rocks. The color distribution of the mineral seems significant. The presence of yellow to brown in the amphiboles is characteristic of the olivine-bearing rocks or of pyroxenite, *i.e.* of rocks high in magnesium; the green color in the amphiboles is typical of norite, *i.e.* of rocks of lower magnesium content. The optic angle of the amphibole in the norite of block 1 is about 72°–80°, that of the amphibole in pyroxenite is about 88°. According to information given in tables by Winchell and Winchell (1951, p. 433) and Herz (1951, p. 994) the higher optic angle is indicative of a higher magnesium content.

Bluish amphiboles are mostly restricted to the pegmatite aureole. The introduction of alkalies by the pegmatites is evident from comparison of the chemical analyses (Table 11) of norite and amphibolite and from the changes in their feldspar composition (Table 3).

The exact composition of amphiboles cannot be obtained from optical data. Optical data are presented in Table 8. The work of Kunitz

TABLE 8.—OPTICAL DATA ON AMPHIBOLE

Rock	Block	2V	Z∧c	γ′ index
Pyroxenite	3	−88°	19°	..
Norite	1	−72 to 80°	12°	..
Amphibolite	1	−71°	14°	1.669
	1	−74°	15°	1.667
	1	−86°	12°	1.659
	3	−75°	19°	..
	3	−86°	13–½°	1.662
	4	−67–½°	....	1.6785
	4	−84–½°	18°	1.6615
	5	−79–½°	11°	1.667
	5	−65–½°	28°	1.681
	6	−72°	12°	1.672
Amphibolite (outer aureole)	7	+84°	16–½°	..

(1930), Winchell (1945), Sundius (1946), and others indicates that the following factors have a modifying influence on the refractive indices: the Fe+Mn+Ti:Mg ratio, the presence of sesquioxides, and the introduction of sodium. An approximation of the composition of an amphibole from a rock in the pegmatite contact aureole ($\gamma' = 1.679$), in which some pyroxenes have survived, gives a 50:50 ratio of Mg:Fe″+Fe‴+Mn+Ti. This information is based on a table given by Sundius (1946, p. 26). Figure 9 shows that this amphibole has the second highest γ' index and the second lowest optic angle. The other points in this curve (Figure 9) represent amphiboles from aureole rocks in which pyroxene has not survived, and which invariably show the blue color characteristic of the introduction of sodium.

Kunitz (1930, p. 232) notes that the refractive index of hornblende decreases with increasing sodium content. The influence of the replacement of Ca-Al by Na-Si (p. 225) is said to be greater than changes in the iron content. If the refractive indices of the amphiboles are plotted against the albite content of the feldspars occurring in the same aureole rocks (Fig. 10), it becomes apparent that the amphibole with the higher refractive index is accompanied by a feldspar of higher sodium content. This would suggest an increase in refractive index

with increasing sodium content in the Caribou amphiboles.

Decrease in the refractive index is accompanied by an increase in the optic angle.

The refractive indices of four garnets ranged between 1.779 and 1.781. Six density values of a contact aureole garnet, determined by D. K. Smith, varied between 3.80 and 3.90; the

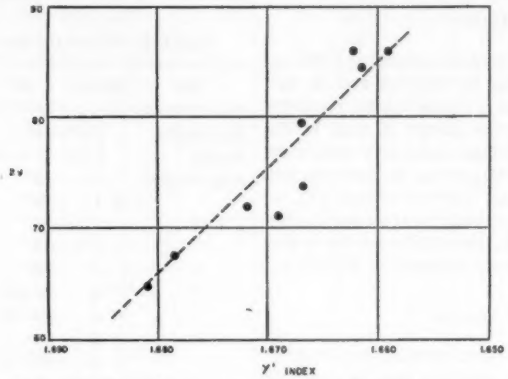

FIGURE 9.—RELATIONSHIP OF γ' INDEX TO 2 V IN AMPHIBOLE

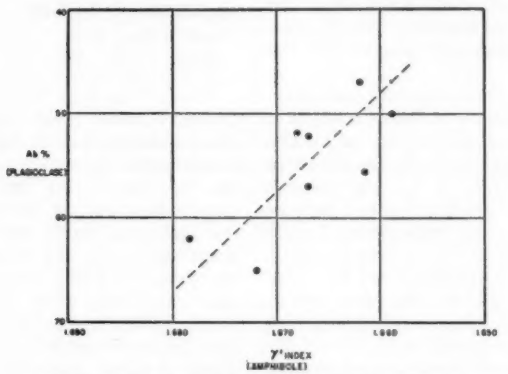

FIGURE 10.—AMPHIBOLE-PLAGIOCLASE RELATIONSHIP

Kunitz (1930) believes that with an increase in the sodium content of the amphibole the optic angle decreases; this observation is here confirmed.

The distribution of the extinction angles of the amphiboles is erratic.

Garnet

Garnet has been found in pyroxenite, olivine norite, norite, rocks of the inner contact aureole, and in the granite pegmatites. It occurs in large euhedral crystals in some of the contact rocks, but usually the crystals are small, commonly microscopic.

average was 3.84. The MnO content of this garnet is 1.52 per cent (C.O. Ingamells, analyst), and its refractive index is 1.781. According to a table by Troeger (1952, p. 12) the garnet composition is as follows: andradite 21, pyrope 47, and almandine 32 mol per cent.

COMPARISON WITH OTHER INTRUSIVE BODIES

General Statement

Brief consideration will be given to a few basic intrusive rocks in which the compositional variation of the rock-forming minerals has been determined. Diabases will not be

included in this discussion. The purpose of this comparison is to bring out the similarities and differences in the compositional variation of the primary minerals in other basic intrusive rocks.

exactly equivalent to the hidden (lower) 60 percent of the Skaergaard". The composition of orthopyroxenes and clinopyroxenes of the lowest exposed horizons of the Skaergaard

TABLE 9.—COMPARATIVE SEQUENCE IN THE ROCK-FORMING MINERALS OF SOME BASIC INTRUSIVE COMPLEXES

	Caribou Lake, Ontario	Stillwater, Montana (Buddington, 1943)	Baltimore, Maryland (Herz, 1951)	Bushveld, S. Africa (Buddington, 1943)	Duluth, Minn. (Buddington, 1943)
Upper part or Caribou late dikes	Ab_{51} Of_{36} $Wo_{41}En_{42}Fs_{17}$ (late dikes)	Ab_{39} Of_{30-40} $Wo_{41}En_{46}Fs_{14}$		Ab_{40-55}	
Above central part or Caribou blocks 4–6	Ab_{43-47} Of_{26-34} $Wo_{41}En_{46}Fs_{13}$ (Norite)	Ab_{20-22} Of_{25-30} $Wo_{41}En_{47}Fs_{12}$ Fa_{70}	Ab_{30-36} Ab_{17-22} (Hornblende) Herz's stage 3	Ab_{30-40}	Ab_{28}
Below central part or Caribou blocks 1, 3	Ab_{37-42} Of_{23-25} $Wo_{46}En_{47}Fs_{13}$ (Norite)	Ab_{16-22} Of_{18-22} $Wo_{41}En_{49}Fs_{10}$	Ab_{30-36} Ab_{17-22} Of_{20-23} $Wo_{41}En_{43-48}Fs_{11-16}$ Herz's stage 2	Ab_{30-30}	
Lower part or marginal facies	Ab_{32-35} Of_{12-16} $Wo_{28}En_{54}Fs_{8}$ Fa_{20} (Picrite)	Ab_{14-21} Of_{15} $Wo_{27}En_{58}Fs_{7}$ Fa_{14}	Ab_{14-22} Of_{10-17} $Wo_{29-40}En_{49-56}Fs_{5-11}$ Herz's stage 1	Ab_{15-25} Of_{15}	

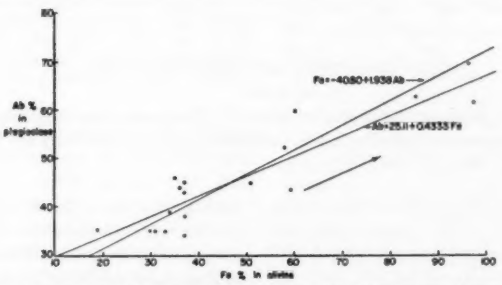

FIGURE 11.—PLAGIOCLASE-OLIVINE RELATIONSHIP IN SKAERGAARD INTRUSIVE BODY
The direction of progressive crystallization is indicated by arrow. Solid lines are linear regression lines. Data from Wager and Deer (1939).

Table 9 gives the comparative sequences of the rock-forming minerals in several basic intrusive complexes.

Skaergaard Intrusive Body

Hess (1940) reviewed Wager and Deer's (1939) classic paper on the Skaergaard intrusive body. He points out that "apparently the exposed portion of the Stillwater is almost

intrusion is practically identical with the most iron-rich pyroxenes of the Caribou Lake and near-by Memesagamesing Lake intrusive bodies and with the pyroxenes from the upper part of the Stillwater. The feldspars of the Caribou and Memesagamesing Lake intrusive bodies, however, are more sodic than those from corresponding positions at Skaergaard and Stillwater.

Figures 11 and 12 show the compositional

variation of coexisting plagioclase and olivine and plagioclase and clinopyroxene in the Skaergaard intrusive body. These figures indicate a linear relationship between the iron relationship between the crystallization of these minerals (Fig. 13, 14). He states: "as orthopyroxene becomes richer in iron, olivine does the same" (p. 986) and "the clino- and

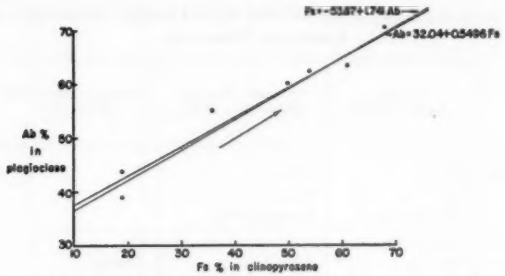

FIGURE 12.—PLAGIOCLASE-CLINOPYROXENE RELATIONSHIP IN THE SKAERGAARD INTRUSIVE BODY
The direction of progressive crystallization is indicated by arrow. Solid lines are linear regression lines. Data from Wager and Deer (1939).

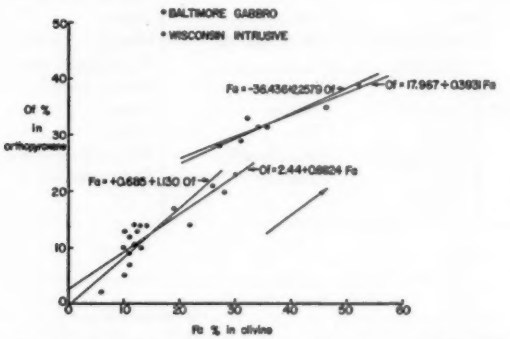

FIGURE 13.—OLIVINE-ORTHOPYROXENE RELATIONSHIP IN BALTIMORE AND WISCONSIN INTRUSIVE BODIES
The direction of progressive crystallization is indicated by arrows. Solid lines are linear regression lines. Data from Herz (1951) and Leighton (1954).

and sodium content. The coefficients of correlation are 0.9166 and 0.9781 respectively, and consequently the correlations are very highly significant. Statistically, about 84 per cent (r^2) of the variation in Fa is explained by the variation in Ab, and about 96 per cent of the variation in Fs is explained by its dependence on Ab.

Baltimore, Maryland, Gabbro

Herz (1951, p. 984, 986, 993) in his study of the Baltimore complex plotted the plagioclase-orthopyroxene, olivine-orthopyroxene, and orthopyroxene-clinopyroxene relationships. His plots of the olivine-orthopyroxene and orthopyroxene-clinopyroxene pairs indicate a definite orthopyroxenes exhibit a straight-line iron enrichment relationship. The clinopyroxene-olivine relation must therefore be similar to the orthopyroxene-olivine" (p. 991–992). The coefficients of correlation and determination for the olivine-orthopyroxene pair are 0.87804 and 77 per cent, respectively, and for orthopyroxene and clinopyroxene are 0.8802 and 77.5 per cent. On Figure 14 the orthopyroxene clinopyroxene relationships of the Baltimore gabbro are compared with those of the Caribou Lake intrusive body. The respective slopes of the regression lines are significantly different. Herz's curve of coexisting plagioclase and orthopyroxene indicates no systematic variation and has not been reproduced in this study.

Gabbro-Granophyre Complex, Northern Wisconsin

Leighton (1954, p. 414–417) discusses the plagioclase, pyroxene, and olivine compositions of the gabbro part of this complex. He states (p. 414, 416) that

> "... a plot of the ferrosilite content of orthopyroxene versus the anorthite content of the plagioclase gives a trend which suggests simultaneous iron and soda enrichment in the orthopyroxene and the plagioclase, respectively. A similar enrichment would be shown by a plot of olivine and plagioclase compositions. The trends would not be sharply defined because of (1) errors in measurement of mineral composition, and (2) failure of the minerals to reach complete equilibrium".

Figure 13 presents the relationship of coexisting orthopyroxenes and olivine in the Wisconsin gabbro as given by Leighton (p. 416). This curve indicates a linear increase in the iron content of orthopyroxene and olivine as crystallization progressed. The coefficients of correlation and determination are 0.94217 and 88.8 per cent, respectively.

On Figure 13 the orthopyroxene-olivine relationships of the Wisconsin intrusive body are compared with those of the Baltimore gabbro. Statistical calculations were made according to the method of Aspin and Welch (Bennett and Franklin, 1954, p. 176–177) for the purpose of comparing the respective slopes of the regression lines for the two intrusive bodies. It was found that the slopes for the Baltimore and Wisconsin data differ significantly.

Comment

The observation that basic intrusive rocks undergo iron and sodium enrichment is now so well established that it requires little comment. Only a small number of basic intrusive rocks have been studied in sufficient detail to permit comparison of the iron or sodium-iron enrichment relationships of mineral pairs from two or more distinct and separate intrusive rocks. The following tentative points seem to emerge, however: 1) A linear regression equation fits the crystallization sequence of mineral pairs of basic intrusive rocks. 2) The slopes of the regression lines for minerals pairs from different intrusive rocks differ significantly.

The points shown on Figures 11 and 12 represent mineral pairs of the Skaergaard intrusive rock sampled at varying distances above the base of the intrusive. They demonstrate the crystallization sequence as traced from the base to the top of the intrusive body. By analogy it can be said that the mineral pairs from the Caribou Lake intrusive body represented on Figures 5 and 6 demonstrate the differentiation trend from the "base" towards the "top" of the intrusive body. This "base" is the narrow constriction at the east end of the intrusive, whereas the "top" is located in the synclinal areas in the western part of the complex. The crystallization sequence at Caribou Lake appears to have progressed westward across the intrusive body. This conclusion is at variance with the structural evidence which points to the presence of funnels in the western part of the intrusive body.

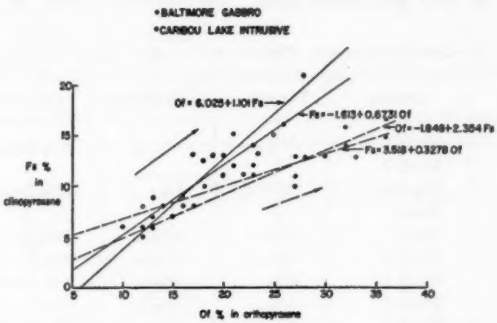

FIGURE 14.—ORTHOPYROXENE-CLINOPYROXENE RELATIONSHIP IN BALTIMORE GABBRO AND CARIBOU LAKE INTRUSIVE ROCK

The direction of progressive crystallization is indicated by arrows. Solid and dashed lines are linear regression lines. Baltimore gabbro data from Herz (1951).

Table 10 presents the differentiation trend of the Caribou Lake intrusive body.

Geochemistry

Major Elements

Wager and Deer (1939, p. 226), Walker and Poldervaart (1949, p. 652) and others have pointed out that the conventional silica-basic oxide diagram is not as satisfactory for demonstrating differentiation trends as a rectangular diagram expressing iron enrichment. In this diagram the $\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3}$ ratio is used as abscissa (Fig. 15). The entire crystallization sequence, on the other hand, is represented on a triangular diagram (Fig. 16) that has as corners MgO, FeO, and ($Na_2O + K_2O$).

Table 10.—Differentiation Trend of Caribou Lake Intrusive Body

Rock	Location	Plagioclase	Orthopyroxene	Clinopyroxene	Olivine
Picrite	Margin Block 2	Ab_{32-35}	Of_{12-16}	$Wo_{38}En_{54}Fs_{8}$	Fa_{20}
Pyroxenite	Margin Block 3	Absent	Of_{18}	$Wo_{38}En_{52}Fs_{10}$	Absent
Norite	Block 1	Ab_{37-40}	Of_{23-27}	Rare or absent	Absent
	Block 3	Ab_{41-42}	Of_{27-28}	$Wo_{40}En_{47}Fs_{13}$	Absent
	Block 4	Ab_{44}	Of_{26-30}	Not determined	Absent
	Block 5	Ab_{43-44}	Of_{30-33}	$Wo_{40}En_{47}Fs_{13}$	Absent
	Block 6	Ab_{47}	Of_{33-34}	$Wo_{41}En_{46}Fs_{13}$	Absent
Late dikes	Blocks 4, 6	Ab_{51}	Of_{36}	$Wo_{41}En_{42}Fs_{17}$	Absent

The chemical analyses used in this study were made on representative samples of the major rock types including picrite, olivine norite, norite, norite with disseminated pyrrhotite, and amphibolite.

The primary differentiates at Caribou Lake have undergone the following changes (Table 11): (1) from picrite to olivine norite. A sharp decrease has taken place in the amounts of MgO and Fe_2O_3, a slight increase in SiO_2, FeO, CaO, Na_2O, and a sharp increase in Al_2O_3; (2) from olivine norite to norite. A continued sharp decrease in the proportions of MgO, a gradual decrease in Fe_2O_3, and FeO, a relatively sharp increase in SiO_2, and a gradual increase in Al_2O_3, CaO and Na_2O.

The postmagmatic changes at Caribou Lake (from norite to amphibolite) include: a continued decrease in MgO, a slight decrease in CaO, FeO, a slight increase in Al_2O_3, TiO_2, and a sharp increase in the alkalies.

The work of Wager and Deer (1939), Walker and Poldervaart (1949), and others has shown that the differentiation of a normal basaltic magma produces iron enrichment throughout a large part of its crystallization history. This is followed in the later stages of crystallization by enrichment in silica and the alkalies. The triangular diagram (Fig. 16), illustrating the differentiation trend of the Caribou Lake intrusive body, shows both iron and alkali enrichment. The line connecting amphibolite and norite in the diagram indicates a trend which turns sharply toward the (K_2O+Na_2O) apex. This line approximately overlaps that of the calc-alkaline series of Walker and Poldervaart (1949). At Caribou Lake this trend has been brought about by postmagmatic changes in the aureole of the granite pegmatites.

The plagioclase-orthopyroxene diagram (Fig. 5) shows that enrichment in sodium and iron takes place with progressive crystallization. The question may be raised whether the net result will be an increase in the absolute concentration of iron or alkalies. The chemical analyses (Table 11) and variation diagram (Fig. 15) indicate that the alkalies increase.[5] FeO increases as the rocks are traced from picrite to olivine norite, decreases from olivine norite to norite, and increases again as the norites are traced westward across the intrusive body. Fe_2O_3 decreases from picrite to olivine norite and norite, and its variation in the norites parallels that of FeO.

The decrease in MgO content in the rock succession picrite–olivine norite–norite is accompanied by increases in alkalies, alumina, and silica. This decrease is brought about by (1)

[5] Analysis 5 (Table 11) is not considered in this discussion as it contains abundant disseminated pyrrhotite, and pyroxenes of anomalous composition.

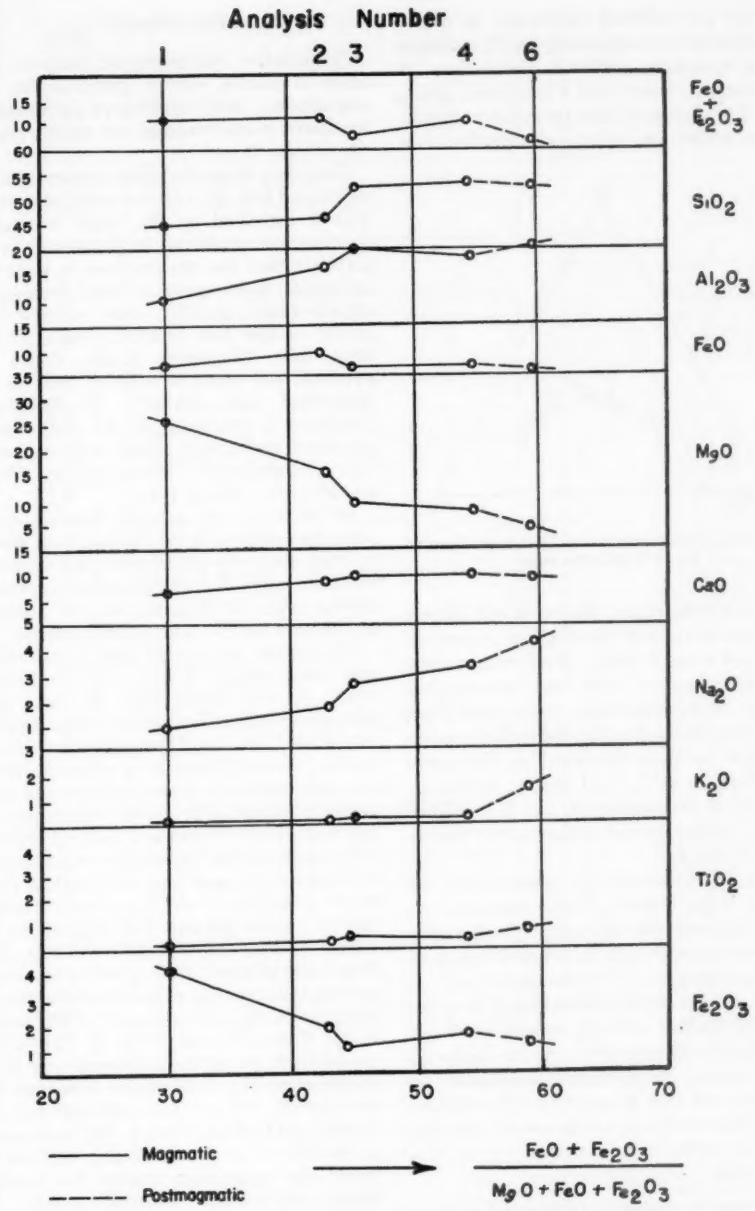

FIGURE 15.—VARIATION DIAGRAM OF CARIBOU LAKE INTRUSIVE ROCK

reduction and ultimate elimination of olivine and serpentine, and apparently by (2) reduction in the pyroxene content.[6] Comparison of samples from blocks 1 and 6 (columns 3 and 4, Table 11)[5] indicates that progressive crystallization within the norite series resulted in a

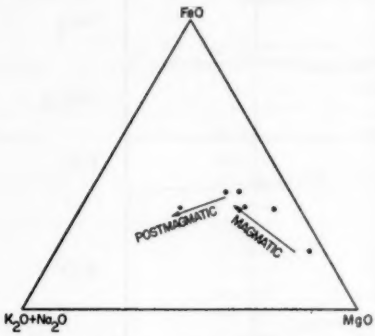

FIGURE 16.—DIFFERENTIATION TREND OF CARIBOU LAKE INTRUSIVE BODY

decrease of magnesium, aluminum and calcium oxides; an increase in alkalies, iron oxides and silica, and a small rise in sulfur content. This observation agrees with the mineralogical changes. In the plagioclases of the norite Na-Si substitutes for Ca-Al and in the orthopyroxenes iron takes the place of magnesium. The almost negligible decrease of CaO, despite sodium enrichment of the plagioclase, can be explained by the increase in the clinopyroxene content (Table 1; Fig. 8).

From this discussion it appears that the Caribou Lake intrusive body manifests two trends in primary differentiation; (1) differentiation involving major changes in modal composition (succession from picrite to olivine norite to norite), and (2) differentiation trends involving chemical changes within a mineral series (*e.g.* plagioclase, orthopyroxene). As the rocks are traced across compositional boundaries (*i.e* from one rock type to another) differentiation trend (1) is dominant; within a given rock type (such as norite) differentiation (2) is more pronounced.

[6] Whereas the decrease and elimination of olivine and serpentine is apparent from thin sections (*see* analyses 1–4, Table 1), the decrease in pyroxene is largely obscured by the presence of reaction rims and symplektite.

Trace Elements

Quantitative spectrographic analyses were made of picrite, olivine norite, norite, and amphibolite, and qualitative analyses of weathered sulfide material and granite pegmatite (Tables 12 and 13).

Chromium shows the highest concentration in picrite and falls off in olivine norite and norite. This is explained by the higher content of ferromagnesian minerals in picrite and by the postulate that the clinopyroxene is a chrome-endiopside. A few grains of spinel, perhaps the chrome-spinel picotite, were observed in picrite. Wager and Mitchell (1950, p. 145) state that "chromium largely enters the pyroxene and there is little or none in the plagioclase and olivine". At Skaergaard chromium is prominent in the early formed pyroxenes and almost absent in those formed later. Cr replaces Fe‴ in the pyroxene (Wager and Mitchell, 1951, p. 164).

Of interest is the gradual decrease in the chromium content of the Caribou Lake norites as these rocks are traced westward across the intrusive body. The low chromium content of olivine norite is puzzling, and no adequate explanation for this observation is offered.

Magnesium and nickel have equal charge and ionic radius. Walker and Poldervaart (1949, p. 655) found that the more basic dolerites are higher in Cr_2O_3 and NiO. Nockolds and Allen (1953, p. 121) state: "that the contents of Cr and Ni tend to be relatively high in the more magnesian parent magmas and to decrease in amount as the parent magma becomes less magnesian". Wager and Mitchell (1950, p. 146) note that nickel is strongly concentrated in the early rocks, and that the residual liquid shows a decline in nickel content. Nickel is said to replace Mg and Fe″ in pyroxene and olivine (Wager and Mitchell 1951, p. 164). Wager and Mitchell (1950, p. 147) attribute the greater abundance of nickel in olivine than in pyroxene to the greater amount of Mg and Fe in the olivine. Fenoglio (1952, p. 121–123) in his study of the peridotite-serpentinites of the Piedmontese Alps finds olivine to contain 0.20 per cent Ni, but pyroxene only 0.05 per cent. Sandell and Goldich (1943, p. 114) were unable to establish a linear relationship between the nickel and magnesium content but found a linear relationship between cobalt and magnesium.

At the Caribou Lake intrusive body nickel shows a gradual decrease from picrite to olivine norite and norite. The nickel content decreases

TABLE 11.—CHEMICAL ANALYSES OF ROCKS

TABLE 11.—CHEMICAL ANALYSES OF ROCKS

	Rock								
	Picrite	Olivine norite	Norite	Norite	Norite with disseminated pyrrhotite	Amphibolite	Picrite (Daly's average)	Olivine norite (Daly's average)	Norite (Daly's average)
Location (Block number)	2	4	1	6	6	5			
SiO_2	42.66	46.24	51.06	52.05	50.66	51.11	41.30	48.78	50.39
Al_2O_3	9.94	16.46	19.51	17.97	17.28	20.70	9.43	18.04	16.06
Fe_2O_3	3.91	2.00	1.09	1.64	1.77	1.17	5.30	1.16	2.43
FeO	6.72	9.25	6.11	7.11	7.45	5.68	8.86	8.94	7.86
MgO	24.89	15.28	8.76	7.26	8.09	4.71	19.94	8.07	8.37
CaO	5.67	7.51	9.24	9.08	10.48	8.89	8.01	8.92	9.20
Na_2O	0.91	1.73	2.50	3.06	2.54	4.21	1.20	2.56	2.61
K_2O	0.22	0.24	.30	.37	0.22	1.29	0.39	0.91	0.79
H_2O +	4.28	0.55	.19	.20	0.22	0.77	4.27	0.69	0.79
H_2O −	0.23	0.08	.10	.09	0.02	0.06			
CO_2	0.06	0.14	.18	.38	0.11	0.03	....	....	....
TiO_2	0.29	0.35	.45	.41	0.20	0.76	0.81	1.48	1.13
P_2O_5	0.03	0.04	.04	.04	0.02	0.12	0.20	0.25	0.20
MnO	0.16	0.16	.12	.15	0.15	0.11	0.29	0.20	0.17
S	0.13	0.14	.10	.45	1.12	0.14	....	....	....
Cl	0.06	0.02	.02	.02	0.03	0.01	....	....	....
less $0 \cong$ S and Cl	100.16	100.19	99.77	100.28	100.36	99.76	100.00	100.00	100.00
	0.07	0.06	.05	.20	0.49	0.06			
	100.09	100.13	99.72	100.08	99.87	99.70	....	....	....
Grain density (gm/cc at temp. indicated):									
	3.077/ 24.8°C.	3.239/ 26.9°C.	2.977/ 25.1°C.	2.981/ 24.5°C.	3.008/ 26.4°C.	2.867/ 29.9°C.			
Analyst	Doris Thaemlitz	Eileen H. Oslund	Doris Thaemlitz	Doris Thaemlitz	Eileen H. Oslund	Doris Thaemlitz			
Remarks	Mode in table 1, no. 1	Mode in table 1, no. 3	Mode in table 1, no. 6			Mode in table 1, no. 19			

in the norites as the rocks are traced westward across the intrusive body. This decrease in the nickel content parallels a reduction in Mg (Fig. 17).

Copper and cobalt are camouflaged by the ferrous ion. They occur in almost equal amounts in the Caribou Lake rocks, and their concentration decreases as the rocks are traced from picrite to olivine norite and norite. In the norites the copper content decreases westward across the intrusive. Inasmuch as the amount of ferrous iron slightly increases, the copper-iron ratio decreases steadily with progressive fractionation.

Patterson (1952, p. 293) notes that "...vanadium shows an initial rise followed by a rapid fall when traced from the silica poor to the silica rich rocks." Nockolds and Allen (1953, p. 121) state that vanadium increases in amount as the parental magma becomes less magnesian. Wager and Mitchell (1950, p. 146) note that the "...vanadium is present in moderate amounts in the early rocks, reaches a maximum about the middle of the series and then falls to very low amounts". They suggest that: "...the total amount of vanadium separating in the early minerals was less than that present in the original liquid. Thus during the early stages of fractionation there was some enrichment of vanadium in the residual liquid". Vanadium replaces Fe''' (Wager and Mitchell 1951, p. 164) in the pyroxene but does not enter olivine (Wager and Mitchell 1951, p. 142). The distribution of V in the Caribou Lake rocks is erratic. It decreases from picrite to olivine norite, increases from olivine norite to norite,

TABLE 12.—QUANTITATIVE SPECTROGRAPHIC ANALYSES

Rock	Block	Boron Per cent	Chromium Per cent	Nickel Per cent	Cobalt Per cent	Copper Per cent	Vanadium Per cent
Igneous							
Picrite	2	.004	.310	.080	.009	.009	.006
Olivine-norite	4	.001	.007	.030	.007	.008	.003
Norite	1	.001	.090	.014	.003	.005	.010
	3	.001	.060	.010	.005	.003	.016
	4	.001	.012	.005	.002	.001	.004
	6	.001	.006	.003	.005	.002	.008
Metasomatic							
Amphibolite	5	.001	.014	.007	.003	.003	.009

Looked for but not found: Sb, As, Cb, Li, Mo, Te, Sn, W, Zr.

D. A. Moddle, Ontario Dept. of Mines, Analyst.

TABLE 13.—QUALITATIVE SPECTROGRAPHIC ANALYSES

Rock	Number 1 Sulfide Concentration "Ore" (weathered)	Number 2 Granite pegmatite
Block	5	5
Ag	T	ND
B	ND	ND
Be	ND	T
Cr	T	ND
Co	T-L	ND
Cu	T-L	T
Mn	L	T
Ni	T-L	ND
Pb	T	T
Ti	L-M	T
V	T	ND
Zn	T	ND

Key to amounts:

L-M—low medium—0.5 to 5.0 per cent

L—low—0.1 to 1.0 per cent

T-L—trace-low—0.05 to 0.50 per cent

T—trace—less than 0.10 per cent

ND—none detected

Looked for but not found in number 1: Sb, As, Bi, Cd, Ce, Cb, In, Hg, Mo, Ta, Te, Th, Sn, W, U, Zr.

Looked for but not found in number 2: Sb, As, Bi, Cd, Ce, Cb, In, Li, Hg, Mo, Ta, Te, Th, Sn, W, U, Zr, Dy.

D. A. Moodle, Ontario Dept. of Mines, Analyst.

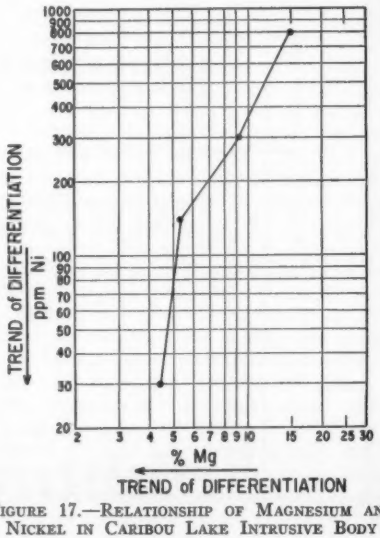

FIGURE 17.—RELATIONSHIP OF MAGNESIUM AND NICKEL IN CARIBOU LAKE INTRUSIVE BODY

and then continues to increase in the norite before sharply decreasing and again increasing as the norites are traced westward across the intrusive body.

At Caribou Lake 0.004 per cent boron has been found in picrite. Olivine norite and norite contain 0.001 per cent boron. Rankama and Sahama (1949, p. 486) contend that boron is an element characteristic of the late stages of magmatic crystallization. In the rocks of the basement complex of southern Lapland (Sahama 1945, as quoted by Rankama and Sahama p. 486) the boron content decreases from the ultrabasics to the gabbros and is least in the granites. The relatively large amount of boron is said to have been introduced during serpentinization. The same probably holds at Caribou Lake, where the picrite has undergone serpentinization, but where serpentine is rare in olivine norite and has not been observed in norite.

Rubidium analyses by X-ray fluorescence (John Hower, analyst) indicated 48 ppm in amphibolite and less than 10 ppm in norite (below the sensitivity of the instrument), or a minimum fivefold increase in rubidium content in the pegmatite contact aureole. This increase is accompanied by an approximately fourfold increase in the K_2O content (Table 11).

The behavior of the trace elements in the course of the fractionation process at Caribou Lake conforms to the observations made on similarly differentiated rocks in other parts of the world. The trace elements chromium, copper, and nickel follow a pattern which suggests that the sequence of crystallization progressed westward across the intrusive body.

Geophysical Surveys

Airborne Surveys

Procedure.—In October 1954, Aeromagnetic Surveys Limited carried out an airborne magnetic and electromagnetic survey of the Caribou Lake intrusive body and surrounding area at the instigation of the Newkirk Mining Company. In a southeasterly direction 194 miles were traversed (Pl. 4). The lines of traverse were spaced at 660-foot intervals. Field data were subsequently plotted on a planimetric base map traced from maps supplied by the Ontario Department of Lands and Forests, Forest Resources Inventory Branch.

Interpretation.—An anomalous magnetic high of strong gradient is defined at the narrow constriction in the eastern part of the intrusive body. This anomaly has not been completely surveyed. However, an aeromagnetic map belonging to the International Nickel Company shows closure of the anomaly as indicated by dots in Plate 4.

Three alternative interpretations will be considered to explain this anomaly: (1) the presence of an oxide or sulfide ("ore") body at depth; (2) the presence of rocks of higher magnetic susceptibility; and (3) the location of a thicker column of rock (norite) of the same magnetic susceptibility.

Alternative 1 is tentatively eliminated as the magnetic intensity is not high enough to indicate the presence of ore, and surface prospecting shows no evidence of mineralization. The magnetic susceptibility of a suite of five norite samples representing outcrops from the eastern extension of the intrusive body to the synclinal areas in the west varies from 54 to 102 $K \times 10^6$ c.g.s. According to these values the anomaly cannot be attributed to the presence of rocks of higher magnetic susceptibility. Elimination of these alternatives leaves the possibility that a thick column of norite is located at the eastern tip of the intrusive.

The pattern of magnetic highs and lows in the northern and southern synclinal areas is complex, and no attempt at interpretation is made.

No conclusions have been reached on the significance of the electromagnetic anomalies covering the northern synclinal area. Anomalies located in Round, Burnt, and Caribou Lake may have been caused by topographic effects.

Linear magnetic highs parallel fault zones in the granite-gneisses, *e.g.* the Wolf River fault and the Caribou Lake fault east of the intrusive body (Pl. 4). Similar observations have been made in the Memesagamesing Lake norite tract (Friedman, 1955, p. 593–595).

Two types of granite-gneiss underlie the area; gneisses giving rise to high magnetic readings, and gneisses producing magnetic lows. Magnetic-susceptibility determinations of gneisses were not made in this study. It is apparent however, that the gneisses with low magnetic readings are confined to the southeastern part of the area mapped.

Ground Surveys

Electromagnetic, resistivity, and self-potential surveys were conducted over the northern synclinal area. This work was followed by a limited amount of magnetometer surveying over the most significant anomalies.

Eight anomalies were outlined in these surveys. All methods used indicated anomalies which coincided with each other. These anom-

alies are linear and have a predominant southwesterly trend. Drilling of these anomalies showed the presence of local disseminated sulfides and graphite. The anomalies are believed to reflect tension joints.

The ground geophysical surveys are not reproduced in this study.

FIGURE 18.—CROSS SECTION SHOWING PROPOSED SUBSURFACE STRUCTURE OF CARIBOU LAKE INTRUSIVE BODY
Along line A–A' (Fig. 4).

SYNTHESIS OF STRUCTURAL, PETROLOGICAL, GEOCHEMICAL AND GEOPHYSICAL DATA

If the two synclinal areas represent channels through which magma ascended, then it should be possible to find some correlation between structure and the petrographic, mineralogical, and geochemical data. A concentric arrangement of rock types around the cores of the funnels might be expected. As the rocks are traced from the centers of the funnels to the margin of the intrusive body, it would seem logical to find certain gradients such as an increase in the Fe/Mg ratio of the pyroxenes or in the Na-Si/Ca-Al ratio of the feldspars or a progressive decrease in some of the trace elements, such as Ni, Cu, or Cr. A comparison of structure with petrology fails to show any such correlation. In fact, a comparison between the differentiation pattern of the mineral series (plagioclase and pyroxene) of the Caribou Lake intrusive body and the Skaergaard and other intrusive bodies suggests that the sequence of crystallization at Caribou Lake progressed from the eastern constriction of the intrusive body to the synclinal areas in the west. This conclusion is at variance with the structural evidence.

The petrologic evidence points to a differentiation path that can be traced westward across the intrusive body and shows no relationship to the synclinal areas. This evidence is briefly summed up:

(1) Ultrabasic rocks are found only near the eastern tip of the intrusive body. No ultrabasic rocks occur in the synclinal areas.

(2) The hypersthene content of the norite decreases and augite increases westward across the complex. This observation is in line with Buddington's (1943, p. 123) contention that " . . . in general, from near the base (of the intrusive) upwards . . . the hypersthenes diminish and augite increases".

(3) The Fe/Mg ratio of the hypersthene in the norite increases progressively westward across the intrusive body.

(4) The feldspar in the norite shows a progressive enrichment in the albite molecule (increase in the Na-Si/Ca-Al ratio) westward across the intrusive body.

(5) Comparison of a chemical analysis of norite from the eastern tip of the intrusive body with an analysis from the southern synclinal area indicates a slight decrease in MgO content to the west and an increase in FeO, Fe_2O_3, and alkalies.

(6) The nickel, copper, and chromium content of the norite decreases westward across the complex.

It was previously noted that the most prominent set of tension joints has a northerly to northeasterly strike. Joints of this type are formed at right angles to the flow of the magma (Balk 1948, p. 31). If this criterion is valid it would suggest that the source of the magma was southeast or northwest of the northern synclinal area. No evidence indicating a northwest source is available, however, petrologic evidence points to a differentiation path originating in the extreme southeast.

The aeromagnetic map offers no decisive criteria that can be employed in interpreting the subsurface structures in the synclinal areas. It is tentatively concluded, however, from the aeromagnetic study in conjunction with magnetic susceptibility determinations that a thick column of norite is located at the eastern tip of the intrusive body. This conclusion confirms the petrological and geochemical evidence and leads to the interpretation that the location of the conduit is at the narrow constriction in the eastern part of the intrusive body (Fig. 18).

Steenland and Woollard (1952) made a geophysical investigation of the Cortlandt Complex, New York, and confirmed two of the funnels proposed by Balk (1927) but rejected a

third. Balk's interpretation of the shape of the Cortlandt Complex was based on structural data. The conclusion arrived at in the present study and supported by evidence from the geophysical investigation at Cortlandt leads to the observation that structural data alone are inadequate to determine the shape of basic intrusive bodies.

The significance of the two synclinal areas in the western part of the Caribou Lake intrusive body is a matter of conjecture. Steenland and Woollard (1952) were faced with the same problem and postulated (p. 1091) that the synclinal area near the center of the Cortlandt complex (Balk's central funnel) "was emplaced downward". They contended (p. 1091) that this could have resulted from:

"a) Subsidence during emplacement resulting from the removal of magmatic material at depth,
b) Subsidence during emplacement resulting from regional tectonic forces (Bucher, 1948), or,
c) Gravity filling of pre-existent surface depressions by extruded material."

Similar interpretations can be advanced to account for the synclinal structures in the western part of the Caribou Lake intrusive body.

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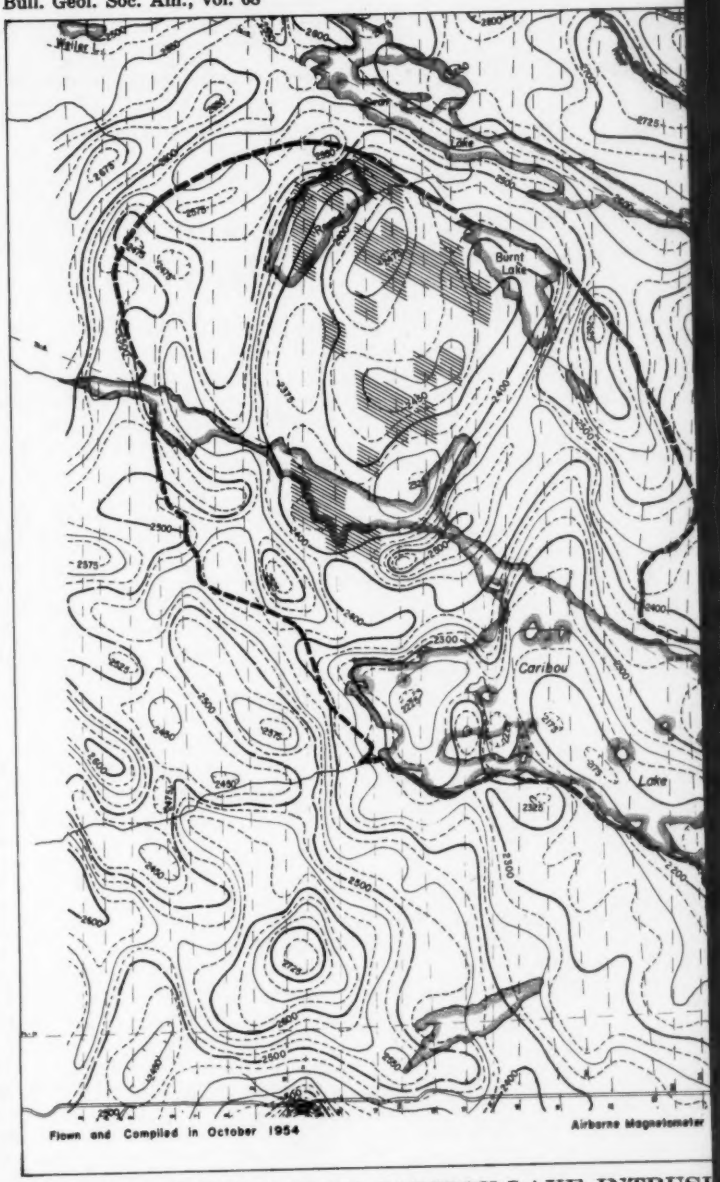
Manuscript Received by the Secretary of the Society, Dec. 17, 1956



AEROMAGNETIC MAP OF CARIBOU LAKE INTRUSI

LEGEND

CONTOUR INTERVAL.....25 GAMMA

MEAN FLIGHT LINE SPACING.....660 FEET

MEAN TERRAIN CLEARANCE.....400 FEET

500 GAMMA CONTOUR.....
100 GAMMA CONTOUR.....
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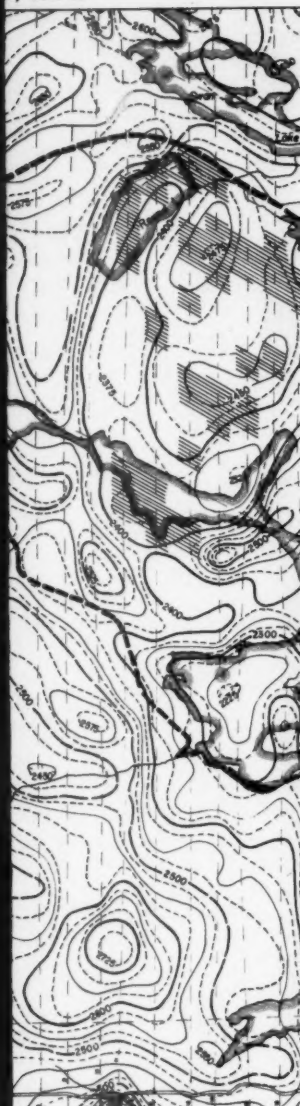
MAGNETIC LOW.....
E. M. ANOMALY.....
FLIGHT LINES.....
OUTLINE OF CARIBOU LAKE INTRUSIVE.....



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TIC MAP OF CARIBOU

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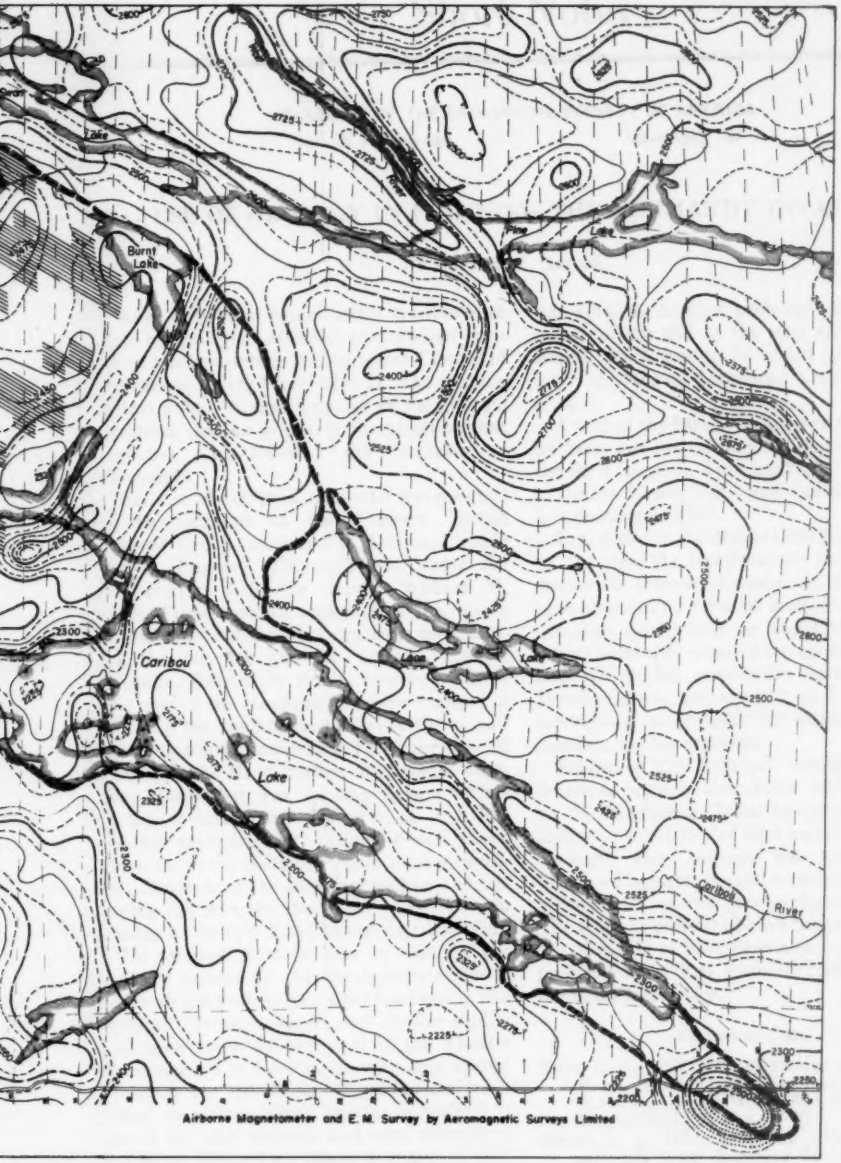

CARIBOU LAKE INTRUSIVE BODY AND SURROUNDING AREA

LEGEND

500 GAMMA CONTOUR
100 GAMMA CONTOUR
50 GAMMA CONTOUR
25 GAMMA CONTOUR
MAGNETIC LOW
E. M. ANOMALY
FLIGHT LINES
OUTLINE OF CARIBOU LAKE INTRUSIVE

SCALE
1000 0 1000 2000 3000 4000 5000
FEET FEET

Short Notes

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1565–1566 NOVEMBER 1957

USE OF GEOLOGY IN PLANNING THE NORMANDY INVASION

BY CHARLES T. SNYDER

Kaye (1957) and Simon (1957) describe work done by geologists attached to the allied command headquarters in the European Theater of Operations during World War II. The purpose of this note is to describe some of the geologic problems that arose during the planning that preceded and followed the Invasion of Normandy.

Early in March 1944, the writer was assigned to Headquarters of the Engineer Special Brigade Group, stationed in Torquay, South Devon, England. This unit was under the command of Brigadier General William M. Hoge. From outward appearances this was a normal administrative headquarters commanding the 1st, 5th and 6th Special Engineer Brigades. Behind the scenes, however, officers and men of the Headquarters were engaged in planning the amphibious phases of the assault on the European mainland. From the time of arrival at the Headquarters until sometime in May, the writer was the only person with a geological background. However, at that time another geologist, Sgt. Samuel Beach Jones, III, came to the Headquarters in another capacity. Throughout the period preceding and following the invasion the writer was N.C.O. in Charge of the G-2 (Intelligence) drafting office.

No attempt is made here to describe the military mission of the Headquarters nor to relate all the geologic problems involved in this mission. Terrain analyses and similar studies had been made by personnel of the Theater Headquarters in London, leaving only special problems for the Group Headquarters level. These special problems usually originated in one of the staff sections and were referred to the writer through the officer in charge of G-2.

Detailed problems connected with interpretation of terrain from aerial photographs arose from time to time. In one such case in which Sgt. Jones collaborated, the suitability of a strip of Omaha Beach* for road construction was studied. Another problem related to the nature of the cliff at the west end of Omaha Beach, which was to be scaled by an assault force of Rangers (American Commando-type troops).

A map of the harbor facilities of Grandcamples-bains was prepared at the request of one of the Staff Sections. This small tidal basin, west of Omaha Beach, was being considered for use in unloading supplies.

In addition to the detailed intelligence library on the area, the Headquarters had a geologic map of the proposed invasion beaches that had been prepared by a British geologist. On several occasions questions arose about the beach or inland areas that required interpretation of this map. In studies undertaken during the pre-Invasion phase, geology was used as a tool of intelligence to evaluate the conditions to be faced during the landings.

Following the Invasion several geological problems arose in connection with operations on the beachhead. These involved the availability or suitability of road material and road drainage. Here geology was used by the Military Engineers in road construction for the maintenance of the communication network. The writer is pleased to have been a part of the Invasion and to have been able to assist when problems arose that required geological interpretation and advice.

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318 C STREET, SALT LAKE CITY, UTAH

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JULY 5, 1957

* A code name used by the Army to protect the identity of the American Sector of the proposed invasion beach. Omaha Beach is on the north coast of France in the province of Normandy.

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1567-1568 NOVEMBER 1957

ADDITIONAL NOTES ON USE OF GEOLOGISTS IN THE EUROPEAN THEATER OF OPERATIONS DURING WORLD WAR II

BY LOUIS J. SIMON

In a recent paper, *Military Geology in the United States Sector of the European Theater of Operations During World War II*, Clifford A. Kaye (1957) listed two capacities in which geologists were able to apply their technical training in the European Theater of Operations (ETO). He showed that the U. S. Army used geologists to make staff studies at the level of Theater Headquarters and as officer personnel in a water-supply unit. The purpose of the present paper is to direct attention to the use of geologists in photographic interpretation.

A number of excellent contributions to the literature have covered the work of photographic interpretation; the *Symposium: Military Photographic Interpretation*, edited by C. G. Coleman and A. C. Lundaal (1948) is especially informative. This collection of papers, like most of the others, mainly reports on general terrain studies obtained by geologic interpretation of the topography.

However, geologists were used during World War II in photographic interpretation of industrial problems. The British apparently first realized the advantages of using interpreters who were specialists in the particular industry to be studied. As a result, the British Intelligence Center located at R.A.F. Station, Medmenham, Bucks, included four geologists in the Industry Section. Later this became the Allied Central Intelligence Unit. This center gathered and disseminated to all branches of the service in the European theater information derived from aerial photographs.

The Industry Section included a so-called P.O.L. (Petrol-Oil and Lubricants) unit. Three British and one American geologist, all former oil company employees, worked in this unit throughout the war. Daily studies covered the location, description, plans, and activities of refineries, oil-storage depots, synthetic oil plants, pipe lines, and oil fields. In addition the unit was frequently called upon for special assignments. For example, one such study resulted in the discovery and mapping of a belt of outcrops of oil shale in southwestern Germany. These outcrops were being mined and processed to obtain crude oil.

In addition to the studies where geology was applied as an aid in industrial interpreting, at least one geologist was a member of the Combined Strategic Target Committee, whose function it was to select bombing targets. This group met weekly, either at the Air Ministry or Fuels Office of the United States Embassy in London, to pick out the strategic targets and set up the weekly priorities. Finally the British War Office in London had a geologist in charge of their P.O.L. Section. He also was a member of the Combines Strategic Target Committee.

Thus the field of Photographic Interpretation provides a definite need for geologists in the service, both in terrain interpretations and in various strategic problems. Here geologic training and background can be utilized to the fullest in determining various industrial and engineering problems. Thus, while geology as a tactical aid in the E.T.O. was reduced to a minimum it was a contributing factor in solving strategic problems. As Mr. Kaye has so ably pointed out, geology is especially useful in the field of Intelligence. Experience suggests that the British appreciate this fact more than we did.

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THE TEXAS CO., 939 S. BROADWAY, LOS ANGELES 15, CALIF.

MANUSCRIPT RECEIVED BY THE SECRETARY OF THE SOCIETY, JULY 5, 1957

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1569-1576, 2 PLS. NOVEMBER 1957

SUPPOSED PERMIAN TILLITES IN NORTHERN MEXICO ARE SUBMARINE SLIDE DEPOSITS

BY NORMAN D. NEWELL

William E. Humphrey (1955) calls attention to poorly stratified, unsorted conglomerates of Permian age in the Valle de Las Delicias about 65 miles northeast of Torreón, Mexico. The succession in which the conglomerates lie is noteworthy as one of the most fully representative and best-documented sequences of Permian rocks in North America. Consequently any considered interpretation about the origin of the conglomerates is likely to be of general interest.

The Las Delicias area, which lies 300 miles south of the well-known Permian deposits of the Delaware Basin of Texas, has been described in a monograph by R. E. King *et al.* (1944) on the Permian stratigraphy and fossil faunas. King (1934) and Kelley (1936) were both impressed much earlier by unusual lithologic characteristics of conglomerates in the lower (pre-*Perrinites*) part of the sequence, and Kelley suggested that they might be glacial tillites. King (King *et al.*, 1944), however believed that the conglomerates must have had some other, possibly tectonic, origin. He cited evidence that near-by Texas enjoyed a warm climate throughout the Permian period; hence he thought that northern Mexico could not have been glaciated at this time. Newell *et al.* (1953) have further amplified evidence for tropical conditions in this part of North America during Permian time.

Humphrey (1955) believes with Kelley (1936) that the conglomerates are glacial deposits, probably ice-rafted to the site of deposition, or even possibly carried by floating islands of vegetation. He cites not only the particular characteristics of the boulder clays but also varvelike characteristics of associated banded shales and siltstones. An alternative view that the poorly stratified and unsorted boulder clays might have been deposited by turbidty currents is considered by him but dismissed without much analysis. In his paper he does not compare closely similar varvelike shales and unsorted boulder beds of the Delaware Basin, which have been explained as deposits of deep stagnant tropical waters, the coarsest fractions of which were believed to have been transported by submarine sliding of unconsolidated sediments down basin slopes (Newell *et al.*, 1953).

During 10 days in August 1956 the writer returned to the Las Delicias area with William M. Furnish; this was our second visit, after an interval of 20 years. Three university graduate student assistants (David L. Clarke, University of Iowa; James X. Corgan and Stanley A. Kling, Columbia University) and G. Robert Adlington, of the American Museum aided in the field work. It is a pleasure to acknowledge the cordial co-operation of Ing. Teodoro Díaz, of Petróleos Mexicanos, and Ing. Guillermo P. Salas, Director of the Instituto de Geología, who gave helpful counsel in planning the field work, loaned maps, and facilitated transportation of fossil collections to the United States for study. After completion of the study, duplicate specimens will be deposited with the Instituto de Geología, in Mexico City.

The primary purpose of the field work was to obtain a comprehensive representation of the fossil faunas (chiefly fusulines and ammonoids). Furnish and Clarke devoted most of their attention to collecting ammonoids, of the black-shale phase. The four other members of the group collected fusulines and other fossils of the limestone phase and documented the lithologic character of the rocks. This required an examination of the entire sequence and cursory inspection of much of the area of outcrop. Time was too short for study of all the significant exposures. Because of Humphrey's important observations and conclusions, particular attention was given to lithologic characteristics of the Permian rocks. A section was measured along the Arroyo La Difunta, with supplementary observations along the adjacent Arroyo La Colorada. These canyons cut directly across the strike of the dipping strata and provide essentially continuous exposure of a large part of the Permian sequence, including the outcrops noted by Humphrey.

The Arroyo Difunta traverse lies parallel with, and about 2–3 km south of, King's "Difunta flank" section (King *et al.*, 1944, Pl. 3, A-A′; p. 13). The results differ in detail from King's in regard to thickness and character of the rocks, but the differences probably are

attributable to lateral variation of the rock units, and the routes of the two traverses do not coincide. Furthermore, field methods differed from those of King; measurements were made with jacob staff, range finder, and compass, whereas King calculated stratigraphic intervals from measurements of dip along a paced traverse. The following compares the two sections.

Difunta flank section, greatly generalized (After King *et al.*, (1944)		Arroyo Difunta section described herein	
Top	Feet		Feet
Bed		Bed	
1. Volcanic rocks	1080	59–64.	795
2. *Kingoceras* shale	220	57–58.	440
3. Volcanic rocks	675	44–56.	1009
4–7. *Timorites* shale	558	43.	400
8–24. *Waagenoceras* shale and volcanic rocks	6398	1–42.	5529+
25–26. *Perrinites* shale	710		
27–33. Conglomerate and volcanic rocks	1203		
Total	10,844	Total	8173

(Lower beds equivalent to King's 25–33 concealed by Cretaceous overlap).

Humphrey's observations especially call attention to the coarse clastics which were overlooked or not sufficiently stressed by previous writers. The writer was unprepared, however, for the discovery that these rocks are quantitatively dominant among rock types of the Las Delicias Permian. Many of the units here cited as conglomerate are more accurately termed graywackes or clay shales, in which pebbles, cobbles, and boulders of extraneous rocks are dispersed. Locally, the coarse fragments constitute as much as one-third of the total rock. More commonly, however, the coarser fraction of the rock forms only a small part of the total mass (Pl. 1, fig. 1). Probably for this reason King, Humphrey, and others have underestimated the quantitative importance of this rock type. As indicated in the section described below, the pebbled and conglomeratic rocks are invariably characterized by poor stratification and poor sorting. Graded bedding is rare.

The matrix characteristically consists of (1) fine-grained graywacke sandstone, mainly pyroclastic, (2) drab to dark-grey, massive mudstone. In the upper part of the section (post-*Timorites* beds) the matrix is reddish chocolate-colored. In the middle part of the sequence, the matrix of a few thin beds of conglomerate consists of silty limestone.

The coarser fragments of the boulder beds consist of a variety of rocks, but the majority are fine-grained igneous rock of intermediate composition. Many are so badly altered that it is difficult to determine their original nature, and a conclusion must await laboratory studies. The igneous fragments range from well rounded to angular. The well-rounded examples presumably were shaped by waves or currents in very shallow water. The sizes represented include granules to a few masses 30–40 feet across, but the majority do not exceed the size of cobbles. None shows faceting or striations comparable to those produced by glaciation.

King *et al.* (1944) cite pillow lavas and several thick sills in the Las Delicias Permian rocks, neither of which was recognized along Arroyo Difunta. Probably equivalent rocks are here classed as graywacke or tuff. Generally, the igneous rocks are much altered so that field distinction among sills, flows, and pyroclastics would be difficult. In most places angular fragments unlike the surrounding matrix are visible in the rocks, but, of course, fragmental material may also be common in sills, flows, and tuffs. Spheroidal weathering is fairly common in some of the graywackes, and this may have been taken for the pillow structure of submarine lava flows. Reworked concretions, broken and tilted at various attitudes, are found at a few horizons where they may have been concentrated by turbidity currents. These generally are associated with pebbles of igneous rocks.

By far the most striking exotics in the conglomerate beds are limestone blocks, some of which are a few hundred feet across (Pl. 1, figs. 1, 2). Many of these larger masses of limestone form the protective caps of conical hills (*e.g.*, King *et al.*, 1944, Pl. 8) scattered picturesquely over the landscape. Limestone blocks are much more characteristic of some horizons than others. Blocks of reef limestone 6 feet or more in diameter occur at many horizons in the interval from bed 9 to bed 44, inclusive. Above bed 44 limestone fragments are

PLATE 1.—PERMIAN ROCKS, ARROYO LA DIFUNTA

FIGURE 1.—Unstratified "boulder clay" in bed 26. Angular fragments include igneous rocks, limestone, and black shale (at right of hammer).

FIGURE 2.—Primary slump fold in conglomeratic beds overlying bedded graywacke, bed 21.

FIGURE 1

FIGURE 2

PERMIAN ROCKS, ARROYO LA DIFUNTA

FIGURE 1

FIGURE 2

EXOTIC LIMESTONE BLOCKS, ARROYO LA DIFUNTA

comparatively rare, and the coarser material of the conglomerate beds is almost entirely volcanic rocks. Beds older than those included in the section contain exotic blocks of calcarenite lithologically unlike the reef blocks. Kelley (1936) and King *et al.* (1944) describe limestone boulders on the west side of the area in thick conglomerates below the *Perrinites* shale, at Las Uvas and Puerto de Sardinas. Several kilometers to the southeast (Malascachas) discontinuous masses of limestone underlie the *Perrinites* shale. Some of these masses (*e.g.*, bed 21, loc. 39 of King *et al.*, 1944) are several hundred feet across and at least 50 feet thick. The contacts are poorly exposed here, but the field relationships suggest that these also may be transported fragments of once-continuous limestone beds. They are massive calcarenite, hence they resemble bank limestone rather than reef limestone. The limestones along the east side of the Paleozoic area tentatively dated by King as Pennsylvanian (?) (King *et al* 1944, p. 7) were examined in the area about 2 miles west of Cerro El Fraile. These are continuous beds of calcarenite limestone. At this place they do not have any of the characteristics of reefs.

King *et al.*, (1944, p. 15) noted that some of the *Waagenoceras* limestones are fairly continuous in the southwestern and northwestern parts of the area but break up elsewhere into "mere lumps". King says, (p. 21)

> The massive limestones probably are the remains of reefs. Laterally, the massive limestones grade into nodular limestone cemented by igneous detrital sand. Some of the individual limestone masses in the nodular beds are as much as thirty feet long and fifteen feet thick; the nodules may have resulted from the breakup of the margins of reefs by waves and by submarine lava flows. The nodular limestone generally thins gradually and resolves itself into several thin layers of sandy limestone. . . ."

The writer believes, as suggested by King, that the limestone "lumps" are not *in situ* reefs but are transported blocks.

Probably they were carried from the place of origin into relatively deep waters by intermittent slumping of volcanic deposits on unstable slopes. Most of the limestone fragments of the *Waagenoceras* and *Timorites* zones (especially our beds 9 to 44, inclusive) consist of microcrystalline bluish-gray limestone in which characteristic reef-frame organisms in the growth position are common. Principal fossils recognizable in the field are fistuliporoid Bryozoa, *Prorichthofenia* sp., and several genera of sycon calcareous sponges. The fabric is closely similar to that of the west Texas reefs (Newell, 1955). The limestone fragments range from pebbles to gigantic blocks and from angular to rounded. There can be no question about the fragmental character of pebbles, cobbles, or even boulders, since the form, contact relations, and orientation of reef fossils within the limestone clearly show that they have been transported. For some of the largest masses the exact stratigraphic relationships are equally clear. In some cases, however, there is reasonable uncertainty, where the exposure is not ideal and the details of the reef fabric are not recognizable.

Excellent exposures occur at many places along Arroyo La Difunta. The large masses of limestone do not have the lenticular form of reefs comparable to those that rim the Delaware Basin (Newell *et al.*, 1953). They are foreign masses in a matrix of conglomeratic graywacke and structureless dark clay and they usually are associated with fragmental igneous rocks and various kinds of nonreef limestone, including reworked limestone concretions. In some cases, the reef fabric forms a rude lineation or pseudostratification which lies at a high angle to the stratification of the surrounding rocks. This discordance indicates that the form of the limestone blocks is not that of the original reefs. Might these masses of reef limestone be patch reefs *in situ* which had become established on a foundation of conglomerate? The writer does not think so. The blocks do not rest on a preexisting surface of conglomerate but lie within the conglomerate. There is no tendency for sorting and stratification of the conglomerate in the inter-block areas. Furthermore, the limestone pebbles and cobbles in these areas are not obviously related to the blocks. They are not arranged in size gradients leading to the blocks, and in many cases they include much material lithologically unlike the blocks. The largest limestone blocks characteristically are much larger than the largest igneous masses.

PLATE 2.—EXOTIC LIMESTONE BLOCKS, ARROYO LA DIFUNTA

FIGURE 1.—Masses of reef limestone and igneous cobbles forming lens (bed 25) within bedded graywacke. Wedge at right may be primary slump mass.

FIGURE 2.—Subangular boulders and blocks of reef limestone in graywacke matrix (bed 26).

The meaning of this disparity is not clear; the difference may reflect original difference in resistance and structure of the source materials.

The limestone fragments that lack the characteristic fabric and lithology of reef limestones are calcarenites that might have been derived from collapse of marginal bank deposits and lithified reef-talus material. Mixtures of limestone types similar to these are prevalent in the clastic forereef limestones of the Delaware Basin (Newell *et al.*, 1953).

The reef-derived limestone fragments of the Las Delicias Permian sequence are conspicuous only in the upper part of the *Waagenoceras* zone and in the *Timorites* zone. This part of the section corresponds closely in time with the greatest reef development around the Delaware Basin—that is, Guadalupian time. Lithologically, however, they more closely resemble the gray limestones of patch reefs of Leonardian and early Guadalupian age in west Texas. Thus, they are probably not relics of great barrier reefs.

About one-third of the section along the Arroyo Difunta traverse consists of dark-gray to black platy shale containing ammonoids and, rarely, brachiopods of the *Nudirostra* type. Banded fissile shale of the sort noted by Humphrey is relatively rare at Las Delicias, but in the Delaware Basin it is a common rock type. The characteristics of this rock and the contained fossils suggest deposition in a deep and relatively stagnant basin in which there were rhythmic, perhaps seasonal, fluctuations in the conditions of sedimentation (Newell *et al.*, 1953, p. 48–94). Irregular fragments of black shale are fairly common in the conglomerate beds (Pl. 1, fig. 1).

The remainder of the section, about one-third, consists of graywacke and chocolate-colored shale, both of which may have originated as water-laid volcanic ash. These rocks show poor sorting, and stratification commonly is in rather thick units.

The writer finds no evidence that supports or even encourages a theory of ice rafting for the unsorted, poorly stratified boulder beds of the Las Delicias Permian. In fact, the lithologic character and the stratigraphic and structural relationships of the boulder beds resemble those of similar deposits in the Delaware Basin for which a glacial origin cannot be seriously proposed. These Mexican boulder beds and volcanic rocks most probably are submarine slide deposits that accumulated in a stagnant basin adjacent to active volcanoes fringed by growing reefs. The sequence is relatively free from examples of graded bedding or other indications of action by turbidity currents. On the other hand, many of the characteristics of submarine slides enumerated by Kuenen (1955) are displayed by the deposits at many horizons within the Las Delicias Permian rocks. The exotic blocks are not known to be associated with strongly folded and overthrust rocks, but otherwise they recall facies of exotic blocks of Newfoundland (Schuchert and Dunbar, 1934), the Himalayas (Heim and Gansser, 1939), Sicily (M. P. Marchetti, personal communication), and elsewhere. Submarine slide deposits are much more abundant in the stratigraphic record than are tillites, and stratigraphers are becoming increasingly alert to their significance.

A paleontologic implication is that many, perhaps most, of the benthonic fossils–fusulines and brachiopods—described from limestones in the Las Delicias area (Dunbar and Cloud, *in* King *et al.*, 1944) were almost certainly taken from exotic blocks, whereas the pelagic ammonoids (Miller, *in* King *et al.*, 1944) from the black shales were probably *in situ*. This interpretation was also made with respect to the Delaware Basin fossils in which most of the benthonic forms were demonstrably introduced by sliding from the basin margin, whereas the pelagic forms show little or no evidence of disturbance after deposition (Newell *et al.*, 1953).

Experience has shown that the stratigraphic sequence of fusulines is in fair agreement with that of the ammonoids in both the Delaware Basin and the Las Delicias area. There is no clearly defined paleontological evidence of mixing of older and younger faunas. Nevertheless, the fusulines of the exotic blocks are certainly generally somewhat older than the associated ammonoids of the shaly matrix. Probably the process of submarine slumping did not normally expose or disturb rocks much older than the accumulating sediments. In this connection, it is significant to note that lithified sediments (concretions, calcarenites, angular blocks of shale) were frequently incorporated in the submarine slides and carried to the basin floor.

King *et al.* (1944, p. 2) thought that the volcanoes responsible for the fragmental volcanics of the Las Delicias area lay west of the Valle de Las Delicias. The present rapid reconnaissance did not provide new evidence on the direction from which the limestone blocks were derived, but further detailed mapping and tracing of

boulder clay units might provide clues in the form of rough textural gradients.

Primary folding produced presumably by slumping of unconsolidated sediments is visible in places along Arroyo La Difunta (Pl. 1, fig. 2). Meticulous study of these structures might indicate the general direction of movement of the sediments. With this problem in mind contorted limestones noted by King *et al.* (1934, p. 15–16) should be re-examined.

Arroyo Difunta Section, Valle de Las Delicias

Supplemented by measurements (beds 47–64) in Arroyo La Colorada; measured by Newell, Corgan, and Kling

Explanation:

* Mainly laminated black shale of euxinic phase containing fossiliferous concretionary limestone with pelagic fauna, ammonoids. Probably deposited mainly *in situ*.

** Unstratified, unsorted clastics, clearly transported from elsewhere and redeposited en masse. Probably deposits of submarine slides.

Beds not designated by asterisks are fine-grained rocks, probably mainly of volcanic origin (graywackes, pyroclastics, including possibly some sills and flows) showing little or no stratification. Some of these may be turbidity deposits.

Top		Thickness, feet
	Bed No.	
Post-*Kingoceras* volcanics		
	**64. Black shale and conglomeratic graywacke, incomplete thickness estimated	150+
	**63. Graywacke and interbedded unresistant conglomerates of well rounded, unsorted igneous fragments	250
	**62. Unsorted, olive conglomerate containing well rounded igneous pebbles in graywacke matrix	120
	**61. Resistant graywacke and tuff, including, at the middle of the section, a large boulder of fine grained igneous rock, about 20 ft. across	80
	**60. Conglomerate, hard, with igneous pebbles in graywacke matrix. Fragments well rounded and unsorted	100
	59. Shaly, olive graywacke with igneous pebbles	95
Zone of *Kingoceras*:		
	*58. Black shale with spheroidal concretions and many lenticular beds of concretionary limestone. A few thin beds of graywacke. Some shale beds are massive and contain reworked concretions and occasional pebbles of igneous rock	395
	**57. *Kingoceras* shale. Black shale with limestone nodules and thin lenses of limestone. Many pebbles of igneous rock, varying widely in size. Upper 10 feet conglomerate with graywacke matrix and many cobbles of black shale and reworked nodules.	45
Post-*Timorites* volcanics:		
	**56. Olive boulder-clay containing boulders of igneous rock, up to three feet in diameter incorporated in unlaminated clay matrix. Interbedded black shale and graywacke. A few reworked concretions	200
	55. Chocolate-colored shale below, grades upward into olive volcanic conglomerate	120
	**54. Conglomerate, reddish-chocolate to olive with unsorted volcanic fragments including many boulders 2 feet in diameter. Forms massive beds 10–15 feet thick with thin lenses of reddish-chocolate shale	180
	53. Shale, reddish chocolate	6
	**52. Conglomerate, reddish chocolate, composed of poorly sorted volcanic fragments	20
	51. Reddish-chocolate shale and graywacke	10
	**50. Unsorted, hard, olive to reddish-chocolate conglomerate with well rounded igneous cobbles	12
	49. Shale, reddish-chocolate, with a few beds of unresistant conglomerate composed of well-rounded igneous fragments	83
	48. Graywacke, reddish-chocolate to brown, with pebble beds of volcanic rock. A few thin partings of reddish-chocolate shale	45
	47. Shale, reddish-chocolate	8
	**46. Unsorted boulder conglomerate of well-rounded igneous rocks in maroon graywaeke matrix	15
	45. Reddish-chocolate shale and silty graywacke with a few brown beds. A few cobbles of igneous rock occur in the upper part in a greenish matrix	110
	**44. Conglomerate with thin, dark, concretion-bearing shales. Cobbles of igneous rock abundant in unlaminated clay matrix. Large, rounded masses of igneous rock, as much as 30 feet across, occur near the base. A few limestone boulders, up to 6 feet across, occur near the top. The unit grades upward into graywacke	200
Zone of *Timorites*:		
	*43. Black shale containing the ammonoid *Timorites* in concretions	400
Zone of *Waagenoceras*:		
	**42. Graywacke with limestone slabs and cobbles	8

Top		Thickness, feet
*41.	Black shale with concretions and a few thin graywacke beds. Lower boundary gradational	450
40.	Graywacke, platy, dark olive	35
39.	Massive, olive tuff	45
*38.	Black shale and concretionary limestone	125
**37.	Conglomerate with pebbles of concretionary limestone and igneous rock	4
*36.	Black shale, hard and platy with a few thin beds of graywacke	140
35.	Massive, olive tuff with interbedded black and gray shale. Contains a few pebbles and cobbles of shale and limestone. Platy at top and bottom, with gradational boundaries	575
*34.	Black shale with concretions and interlayered tuffaceous graywacke. Abundant *Nudirostra* found near base	155
**33.	Gray-buff graywacke and minor amounts of interbedded black and gray-banded shale. At base, contains boulders of reef limestone 5 feet across. Numerous cobbles and boulders of igneous rock. Beds locally contorted and brecciated	200
*32.	Black shale with a few hard, thin, concretionary beds	92
31.	Olive-drab graywacke	20
**30.	Gray limestone, arched in a recumbent fold over undisturbed beds below. Associated with transported blocks of reef limestone	10
*29.	Shale, black, with thin interbeds of olive-buff graywacke	105
28.	Olive-drab graywacke with a few thin beds of black shale. Shows spheroidal weathering	45
*27.	Black shale with fossiliferous concretions. A few thin beds of graywacke, sandstone, conglomerate, and reworked phosphatic concretions near the base. Some contorted bedding above basal contact	220
**26.	Bouldery volcanic tuff with both igneous material and reef limestone. Several zones contain large blocks of reef limestone. This unit makes the major "falls" across the Arroyo	450
**25.	Boulder conglomerate with blocks of reef limestone up to 15–20 feet overlying conspicuous crumpling. Limestone blocks form conspicuous ridge north of the Arroyo. Matrix consists of tuffaceous sandstone and black shale	20
**24.	Black shale and conglomeratic, tuffaceous sandstone	60
**23.	Conglomerate. Tan, graywacke matrix and angular pieces of black shale	5
*22.	Black shale	25
**21.	Black shale and conglomerate, tuffaceous sandstone. Pebbles mainly black shale and black limestone concretions. Shows slump folding	15
**20.	Tuffaceous conglomerate and sandstone, drab to tan, filled with reworked, black limestone concretions	27
*19.	Interbedded black shale and olive, tuffaceous sandstone	17
**18.	Conglomerate with igneous and limestone cobbles. Interbeds of drab, tuffaceous sandstone	55
17.	Volcanic tuff, dark greenish-gray with black shale inclusions. Shows spheroidal weathering (Beds 17 to 20, inclusive, produce steep steps across the Arroyo La Difunta.)	10
*16.	Black shale with phosphatic concretions and thin interbeds of tan graywacke	55
**15.	Boulder conglomerate with graywacke matrix, black shale and reef limestone boulders, igneous pebbles, and reworked black concretions (limestone)	22
*14.	Black shale	20
**13.	Boulder conglomerate with masses of reef limestone up to 40 feet by 10 feet	15
**12.	Black shale with scattered boulders of lava, tuff, and limestone	125
**11.	Boulder conglomerate similar to 9c	20
*10.	Black shale	20
**9c.	Conglomerate, simulates bedded limestone: limestone matrix with cobbles of limestone. A few fusulinids. Exposed in Arroyo La Difunta at the site of a cave near Casa Difunta. The upper part grades into black shale northward, within 100 yards. The lower portion contains blocks of reef limestone up to 15 feet across	40
*9b.	Shale, black, with concretions	75
**9a.	Conglomerate, with tuffaceous sandstone matrix containing limestone blocks up to 10 feet across. Some limestone blocks are tan to gray and contain abundant fusulines of several species. Others are typical light-gray reef limestone with abundant sycon sponges and *Prorichthofenia*	15
8.	Sandstone, tuffaceous, buff-colored, and black shale	225
**7.	Conglomerate, fossiliferous, matrix mainly limestone, pebbles small	12
*6.	Black shale	30
**5.	Lens of fusuline limestone, dark gray, exposed on both sides of Arroyo La Difunta	2

Top		Thickness, feet
	*4. Black shale and platy, tuffaceous sandstone. Contains *Waagenoceras* in concretions	250
	**3. Tuffaceous, olive sandstone and unsorted conglomerate with subrounded limestone fragments and reworked black concretions. Cobbles, up to 8–10 inches in diameter, include fine-grained, blue-gray limestone and fine-grained igneous rocks	40
	*2. Black shale with minor interbeds of highly fossiliferous, concretionary, brown limestone. A few beds of olive, tuffaceous sandstone. Rare ammonoids at bottom and top	150
	1. Interbedded fine, olive to light gray, tuffaceous graywacke, locally with cannon-ball concretions. Minor amounts of black, unfossiliferous shale	1,500
Base of section concealed by Cretaceous overlap. Total section exposed:		8,173

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Manuscript Received by the Secretary of the Society, February 14, 1957

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1577–1580, 1 FIG., 1 PL. NOVEMBER 1957

LOWERRE QUARTZITE PROBLEM

BY MATTHEW F. NORTON AND ROSS F. GIESE, JR.

Regional and Geologic Setting

Southern New York State east of the Hudson River consists of three major structural units. In the southernmost, the New York City and Westchester County metamorphic zone, parallel belts of metasedimentary rocks lie in what are usually considered north-northeasterly trending isoclinical folds; this terrane disappears into the subsurface to the west along the line of the Hudson River, except for a small exposure in Hoboken, New Jersey. North of this metamorphic zone, the Hudson Highlands, a faulted massif of Precambrian gneisses and fault slivers of younger rocks, cuts roughly east-west across the trend of the folds to the south; this appears to be traceable into the northeast-trending New Jersey Highlands west of the Hudson River. North of the Hudson Highlands, the north-northeasterly trend of the structures resumes, but here the rocks on the west are less intensely metamorphosed and are fossiliferous in places. According to Balk (1936), this belt may be traced east-southeast through increasing rank of metamorphism.

Previous Work

Throughout the earlier work leading up to the publication of the New York City Folio (Merrill *et al.*, 1902), the Lowerre quartzite was identified as the basal unit of the Manhattan group, a quartzite-limestone-schist sequence lying over the Fordham gneiss in the southernmost metamorphic belt. Berkey (1907) suggested that it might only be an upper quartzitic facies of that gneiss. Recent papers on the geology of Westchester and Putnam counties (Prucha, 1956; Scotford, 1956) have either denied the existence of the Lowerre quartzite or called it, when recognized, a "sheared gneiss."

Type Locality

The writers recently visited the type locality for the Lowerre quartzite on the hill south of Lawrence Avenue in Yonkers and collected specimens of the rock for thin-section analysis. The remaining outcrop (Pl. 1, fig. 1) is exposed part way up the hill, but otherwise the rock is covered. Here, the quartzite when better exposed was described as 5–10 feet thick. In hand specimens it is a thin-bedded white quartzite showing scattered flakes of muscovite along fracture planes; it breaks easily into rhomboid fragments along these joints. Under the microscope (Pl. 1, fig. 2) it shows the following minerals:

	Per cent	
quartz	65–70	many grains strained, some sutured;
plagioclase	3–5	the feldspars generally show
orthoclase	10–15	alteration on the edges,
microcline	3–5	and some are clouded;
muscovite	3–5	all secondary;
allanite	<1	
apatite	<1	
biotite	<1	
pyrite	<1	
zircon	<1	hypidiomorphic to xenomorphic

Although the observed percentage of feldspar relative to that of quartz would place this rock near the arbitrarily defined border between feldspathic sandstone and arkosic sandstone, the fact that some of the quartz is present as cement suggests that the relative percentage of feldspar in the original sediment may have been higher, and that, therefore, prior to metamorphism the Lowerre quartzite may have been locally derived arkosic sandstone. There is no megascopic or microscopic evidence of shearing sufficiently intense to have produced this rock from a gneiss without an intervening period of selective weathering and sedimentation.

Other Outcrops

Figure 1 and Table 1 give the locations of other outcrops of the Lowerre quartzite. In most places the rock is thin-bedded; it is white to gray white, although in many places it is stained brown. In the Crugers outcrop it is grayish and apparently altered by the near-by Cortlandt complex.

At Sparta Landing the quartzite is about 40 feet thick. The major structural feature exposed in a railroad cut south of the landing is a

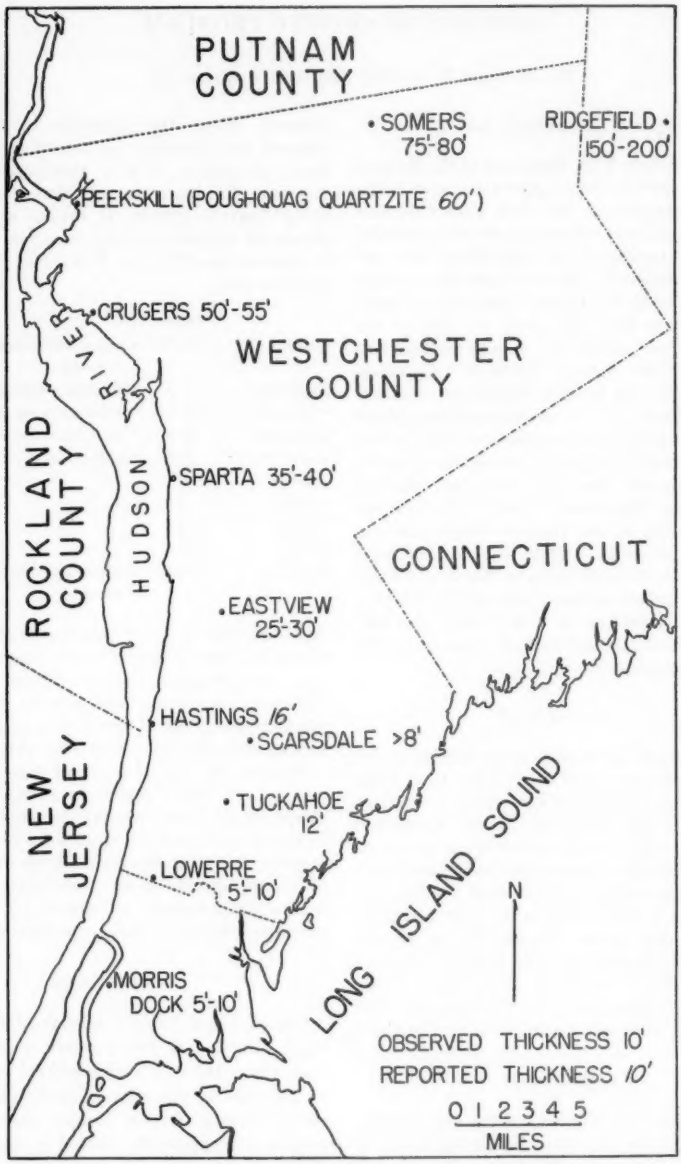

FIGURE 1.—LOCATION OF OUTCROPS OF LOWERRE QUARTZITE

TABLE 1.—LOCATION OF LOWERRE QUARTZITE OUTCROPS

Crugers: On the south side of Maiden Lane in the bed of Furnace Brook below the lowest dam

Eastview: On the west side of the railroad cut 1⅝ miles south of Eastview Station, measured along the tracks of the Putnam Division of the New York Central Railroad

Hastings: On and behind the hanging wall of the quarry on the east side of the road running along the Old Croton Aqueduct

Lowerre Station: About 20 feet south of Lawrence St. in the abandoned railroad cut east of Saratoga St.

Morris Docks: 200 yards south of the Morris Heights Station on the east side of the railroad cut just below the New York Thruway

Peekskill: Poughquag quartzite on the southeast bank of Peekskill Creek

Ridgefield: On the hill west of New Road which runs west from U.S. Route 7, 1.4 miles south of its junction with Conn. Route 35; the quartzite is well exposed in a cliff on the northwest side of the southern part of the hill and in ledges on the southeast side of the central part of the hill, but no contacts are exposed. This outcrop was first located by James W. Clarke.

Scarsdale: 900 yards north of Scarsdale on the west side of the railroad tracks, in a cut on the right-of-way of the Harlem Division of the New York Central Railroad.

Somers: On the hill north of U.S. Route 202, 0.8 miles west of its junction with N. Y. Route 100; the quartzite is here overturned and lies under what is usually identified as "Highlands gneiss". There is a difference of 10–15° in places between the lineations of the gneiss and the bedding of the quartzite.

Sparta: 260 yards south of the old Sparta Landing on the east side of the tracks of the main line of the New York Central Railroad; vertical beds exposed northwest of Sparta Brook

Tuckahoe: In the west wall of the old Norcross Brothers Marble Quarry

recumbent fold. Along the north side of Sparta Brook, part of the arch-bend of the fold is exposed showing almost vertical bedding striking northeast.

The outcrop in the railroad cut north of Scarsdale shows that the gneiss there behaved as a competent mass, while the quartzite behaved incompetently. The quartzite, whose full thickness is not exposed here, is best shown in a small attenuated recumbent fold. The gneiss, which does not show complex folding here, lies above this fold, forming two ledges separated by badly weathered fault zones. In the lower ledge the gneiss dips 49° NW. Primary structures (cross bedding and either cut-and-fill structure or linguoid ripple marks) show it to be right side up. Above this the gneiss in the second ledge dips 62° NW. This outcrop suggests that the concept of the synchronous isoclinal folding of the gneiss and the overlying beds may warrant re-examination.

General Stratigraphic and Structural Relations

Minor stratigraphic features appear in the quartzite in a number of localities. These are best shown in a railroad cut 1⅝ miles south of the Eastview station, as measured along the tracks. In this outcrop there is a 3- to 4-inch incompetent bed about 5 feet below the top of the quartzite. A similar layer is also found at Sparta and at Scarsdale. In all these localities there is also a zone with micaceous, presumably originally argillaceous partings in the middle section of the quartzite. At Eastview this zone begins 10 feet above the base of the formation.

If the quartzite were merely an upper, more quartzose member of the Fordham gneiss, it should exhibit structural behavior similar to other quartzose layers in that gneiss. However, this has not been observed in any outcrop examined.

Conclusions

Although the existence of a quartzite, possibly locally derived, between the Fordham gneiss and the Inwood limestone at a number of points cannot be denied, its stratigraphic significance in what is now called the New York City group (Prucha, 1956) may have been overemphasized. Quartzites, like the sandstones from which they are derived, are notorious for their sudden appearances and disappearances. The absence or apparent absence of a sandstone or quartzite may be a result of nondeposition, of concealment by glacial or alluvial deposits, of removal by erosion or faulting, or of granitization (Dana, 1881) and cannot be taken as disproving correlation as suggested by Prucha (1956). On the other hand, the specific existence of the Lowerre quartzite cannot of itself be considered proof that the New York City group is only a more highly metamorphosed extension of the Cambro-Ordovician rocks north of the Hudson Highlands, despite the similarity of their lithologic sequences. Such a conclusion

would be justified only if it can be shown that the Lowerre is a direct extension of the basal unit of this series, the Poughquag quartzite, or correlative with it on sound stratigraphic ground.

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Manuscript Received by the Secretary of the Society, January 8, 1957

Plate 1.—LOWERRE QUARTZITE

Figure 1.—Outcrop of Lowerre Quartzite at Type Locality
Showing its thin-bedded nature and fluted folding

Figure 2.—Thin Section of Lowerre Quartzite from Type Locality
Magnification ×80, crossed nicols

FIGURE 1

FIGURE 2

LOWERRE QUARTZITE

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 68, PP. 1581-1584, 2 FIGS NOVEMBER 1957

POLYMORPHISM OF MICAS IN THE MINERAL BLUFF AND EPWORTH QUADRANGLES, GEORGIA

BY VERNON J. HURST

Mica-bearing rocks in the Epworth and Mineral Bluff quadrangles, north Georgia (Fig. 1), bear the imprint of at least four geologic environments. Three environments are connected with zones of regional metamorphism: biotite zone, garnet zone, and staurolite zone. Superimposed on the mineral assemblages of these three environments are the effects of post-metamorphic hydrothermal alteration (Hurst, 1955, p. 104).

To determine whether there is any systematic variation in polymorphism correlatable with environment, 97 samples from this area have been examined by powder X-ray-diffraction techniques. Six samples are from the biotite zone, 5 from the garnet zone, and 86 (27 of them showing the effects of post-metamorphic hydrothermal alteration) from the staurolite zone.

The simplest mica polymorphs theoretically possible are 1M, $2M_1$, $2M_2$, 3T, 2O, and 6H[1] (Smith and Yoder, 1956, p. 210–214). Not all these polymorphs have been found in natural micas. Among biotites, the 1M, $2M_1$, and 3T, as well as the more complicated 18M and 8Tc (triclinic), structures have been recognized (Hendricks and Jefferson, 1939, p. 734, 738–746, 756; Amelinckx and Dekeyser, 1953; Smith and Yoder, 1956, p. 216). Among muscovites, the 1M, $2M_1$, 3T, and 12M structures have been described (Yoder and Eugster, 1955, p. 244; Jackson and West, 1931; 1933; Hendricks and Jefferson, 1939, p. 738; Axelrod and Grimaldi, 1949, p. 559–572; Smith and Yoder, 1956, p. 216). The $2M_2$ polymorph is known only among lepidolites (the 6M structure of Hendricks and Jefferson, 1939, p. 747, renamed $2M_2$ by Smith and Yoder, 1956, p. 215). The theoretical polymorphs 2O and 6H have not been found in nature or in synthetic products.

Several efforts have been made to relate the polymorphism of the micas to growth environment. For illites and hydrous micas, "no definite correlation between the type of polymorphic crystallization and either chemical composition or geological occurrence has been observed" (Levinson, 1955, p. 41). The stability ranges of the muscovite polymorphs could not be experimentally fixed by Yoder and Eugster (1955). They did, however, effect the transformation $1M_d$ (disordered) → 1M → $2M_1$. They regard this sequence as probable in the progressive metamorphism of sediment, as the 1M polymorph is obtained experimentally at low temperatures. They report several examples of natural $1M_d$ and 1M muscovites and phlogopites belonging to a low-temperature environment (Yoder and Eugster, 1955, p. 249–251; 1954, p. 182). Smith and Yoder (1956, p. 229) have concluded

> "there are factors other than environmental that determine the type of polymorph. Composition clearly has a strong influence on the structure . . . the variation of structure in the natural micas cannot . . . be attributed solely to changes in pressure and temperature."

In the Mineral Bluff and Epworth quadrangles, the muscovite micas all have the $2M_1$ structure (Fig. 2), and the biotite micas have either the 1M or 3T structures (which are not distinguishable by X-ray powder patterns).

The principal difference in the physical conditions attending the growth of the micas in the three metamorphic environments was probably temperature. The post-metamorphic alteration environment differed, at least, in the amount of available water. Certainly each of the four environments differed in some way from the others, and these differences are not reflected in the polymorphism of the micas. These results support the conclusion reached by Smith and Yoder that factors other than environment may determine the type of polymorph.

Acknowledgment

The X-ray work was done at the Geophysical Laboratory, Washington, D. C., under the supervision of Dr. Hatten S. Yoder, who has edited the manuscript.

[1] The first numeral in each symbol gives the number of mica layers in the unit cell. The letters give the symmetry. M = monoclinic; T = trigonal; O = orthorhombic; H = hexagonal. To distinguish the two 2-layer monoclinic (2M) polymorphs subscripts 1 and 2 are used.

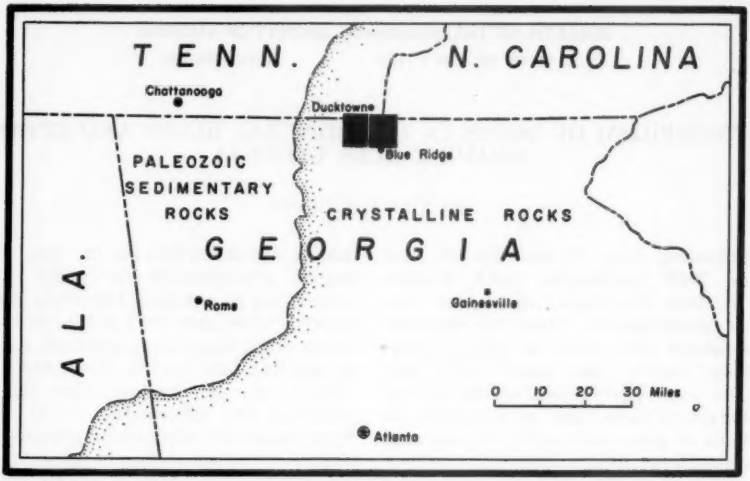

FIGURE 1.—INDEX MAP SHOWING LOCATION OF THE EPWORTH AND MINERAL BLUFF QUADRANGLES

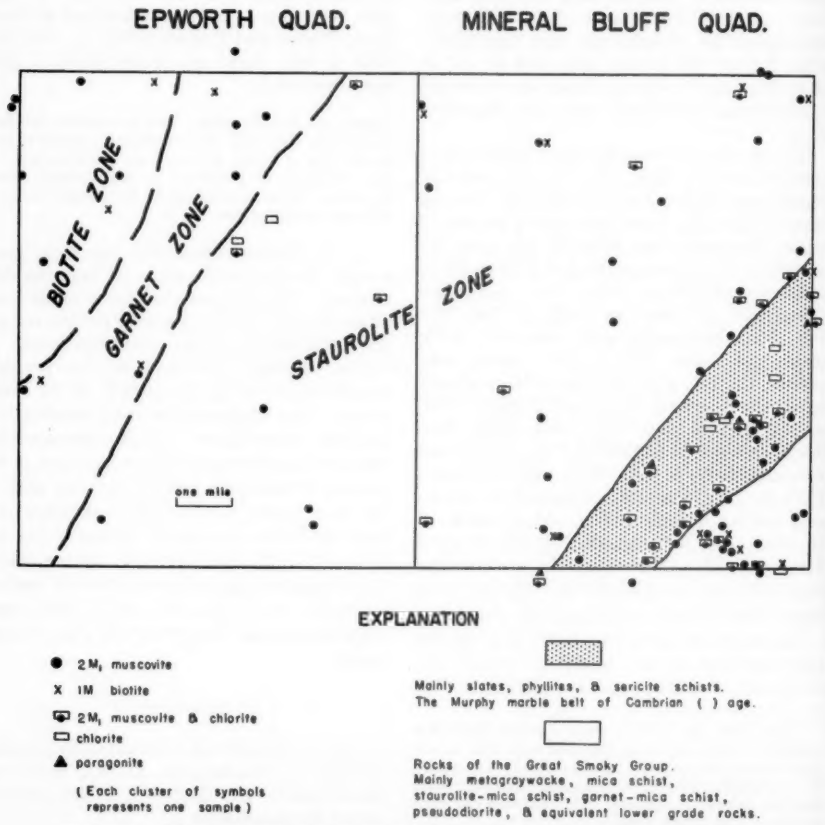
EPWORTH QUAD.
MINERAL BLUFF QUAD.
BIOTITE ZONE
GARNET ZONE
STAUROLITE ZONE
ZONE
one mile
EXPLANATION
2M1 muscovite
1M biotite
2M1 muscovite & chlorite
chlorite
paragonite
(Each cluster of symbols represents one sample)
Mainly slates, phyllites, & sericite schists. The Murphy marble belt of Cambrian ( ) age.
Rocks of the Great Smoky Group. Mainly metagraywacke, mica schist, staurolite-mica schist, garnet-mica schist, pseudodiorite, & equivalent lower grade rocks.

FIGURE 2.—MICAS IN THE EPWORTH AND MINERAL BLUFF QUADRANGLES, GEORGIA

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Manuscript Received by the Secretary of the Society, July 1, 1957.